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CONTENTS

	Pages
Trace-element studies, Santa Rita, New Mexico. By Donald Graf and Paul F. Kerr	1023–1052
Metamorphism and associated mineralization in the Philipsburg region, Montana. By William T. Holser	1053–1090
Submarine geology of Bikini Lagoon as indicated by dispersion of water-borne explosion waves. By Milton B. Dobrin	1091–1118
Pleistocene history of Coastal Alabama. By Charles W. Carlston	1119–1130
Stratigraphy and structure of the Oak Hill succession in Vermont. By Verne H. Booth	1131–1168
Short Notes	
Trends in Geomorphic Research. By Charles W. Carlston	1169–1170
Pollen analysis in Old and New Mexico. By Paul B. Sears	1171–1172

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PAPERS IN PRESS FOR FORTHCOMING ISSUES

BLUE RIDGE FRONT—A FAULT SCARP. By William A. White

GEOLOGY OF THE BROKEN HILL ORE DEPOSIT, BROKEN HILL, N. S. W., AUSTRALIA. By J. K. Gustafson, H. C. Burrell, and M. D. Garretty

TERRACED HIGHWAY SIDE SLOPES IN LOESS, SOUTHWESTERN IOWA. By CHARLES S. GWYNNE

DIABASE OF THE GEORGIA PIEDMONT. By J. C. Lester and A. T. Allen

MEASUREMENT OF PALEOTEMPERATURES AND TEMPERATURES OF THE UPPER CRETACEOUS OF ENGLAND, DENMARK, AND SOUTHEASTERN UNITED STATES. By H. C. Urey, H. A. Lowenstam, S. Epstein, and C. R. McKinney

CARBONATE-WATER ISOTOPIC TEMPERATURE SCALE. By Samuel Epstein, Ralph Buchsbaum, Heinz Lowenstam, and Harold C. Urey

GEOLOGY OF THE EASTERN VENEZUELA BASIN. By Hollis D. Hedberg

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TRACE-ELEMENT STUDIES, SANTA RITA, NEW MEXICO

By Donald L. Graf and Paul F. Kerr

CONTENTS

	Page
Abstract	1023
Introduction	1023
Acknowledgments	1024
Study of trace element distribution	1025
Procedures	1025
Choice of tracer elements	1026
Sampling methods	1026
Spectrographic procedure	1027
Location and general geology	1028
Tertiary area	1031
General statement	1031
Prospecting in the Tertiary	1031
Bleached zones	1033
Pink clay horizons	1033
Spectrographic data, Tertiary surface specimens	1033
Spectrographic data, Tertiary clays	1035
Pre-Tertiary ore bodies	1038
Empire Zinc and Oswaldo Mines	1038
Georgetown and Lone Mountain	1042
Groundhog mine	1043
Spectrographic data, surface limestone specimens	1044
Spectrographic data, underground limestone specimens	1047
Conclusion	1051
References cited	1052

ILLUSTRATIONS

Figure	Page
1. Fused quartz hood used to introduce CO_2 atmosphere	1027
2. The center-pole electrode	1027
3. Density/Intensity curve for Kodak number 33 spectrographic plates	1028
4. Index map, Santa Rita district	1029
5. Stratigraphic section, Santa Rita area	1030
6. Approximate location of Tertiary specimens	1032
7. Data for Tertiary surface specimens	1034
8. Tertiary clay specimens, Section I	1036
9. Tertiary clay specimens, Section II	1036
10. Tertiary clay specimens, Section III	1037
11. Tertiary clay specimens, Section IV	1038
12. Date for Tertiary clay specimens	1039
13. Location map for Georgetown surface specimens	1042
14. Location map for Lone Mountain surface specimens	1043
15. Location of Hanover surface specimens	1045
16. Data for surface limestone specimens 1–60.	1046
17. Data for surface limestone specimens 61–117	1047
18. Specimen location map and section, Oswaldo mine	1048
19. Data for underground limestone specimens 1–57	1049
20. Data for underground limestone specimens 58–109	1050

TABLES

	Page
Table 1. Trace element highs in Tertiary clay specimens	1040

Abstract

Emission spectrography has been used to study distribution of trace elements in Paleozoic limestones surrounding Pb-Zn ore bodies near Santa Rita, New Mexico, and subsequent transportation of these traces into overlying Tertiary beds. Analyses are given for Pb, Zn, Cu, Ag, and V, important ore metals, and Mn and Al, prominent in pre-ore silication.

Pb highs from Tertiary fault zones may be correlated with ore bodies in pre-Tertiary rocks, but Zn has been leached unless considerable gouge is present. Pb and Zn highs from a remote part of the Tertiary area encourage further exploration.

A zoning of traces occurs in Tertiary montmorillonite zones bordering the ore-localizing Groundhog fault. Near a producing mine, Ag was deposited for 20 feet from the fault, and Pb for 40 feet. Most Zn highs are in a 60-foot zone, but a few extend as much as 250 feet.

In limestones around Pb-Zn ore bodies, trace-element distribution is controlled by small fractures, and penetration into massive wall-rock is negligible. Zn highs occur close to ore bodies, and Pb highs farther away. The shape of the Pb gradient seems controlled largely by temperature, and the relative importance of pre-ore silication and ore-deposition may be important here.

Introduction

In the course of this work, spectrographic analyses have been applied both to fault gouge

and bed rock, utilizing specimens from outcrops, diamond drill cores, and mine workings. The elements detected have been relatively undisturbed since original deposition, except for some redistribution by ground water. The pattern of trace elements observed around ore bodies or near mineralized areas and the zones resulting are believed to shed light on the nature and migration of the mineralizing solutions.

A number of factors have facilitated the solution of the problem of analyzing many specimens from numerous confirmatory traverses, needed to ascertain the nature and significance of general trends in spectrographic data. Frequent reference has been made to detailed geologic reports on the Santa Rita area (Lasky, 1936; Schmitt, 1939; Wright and Kerr, 1947; Patterson and Kerr, 1947; Kulp and Kerr, 1947) and unpublished reports in company files, which have reduced in amount the geologic field work required. The emission spectrograph, effective in the range 0.0001%—1.0%, proved sufficiently sensitive and accurate. More time-consuming chemical pre-enrichment of samples, or colorimetric determination of individual elements, were generally omitted. Analysis of representative specimens eliminated many elements from consideration. Concentrations were plotted in relative numbers, based upon the microphotometer scale, and only when germane were they converted to absolute values.

The words "trace" and "trace element" refer to minute quantities. They have no genetic significance, and are not restricted to any group of elements. When considering the usefulness of distribution studies of trace-elements in finding hidden ore bodies, temperatures of deposition, formational contacts, etc., they are called "tracers" or "tracer-elements." The term "dike-fault zone" has been applied to individual sheet-like loci of repeated faulting and dike intrusion which extend southward from the Hanover intrusive. "Metasomatism" involves significant addition of outside material, while "metamorphism" is limited to rearrangement of components of the original rock.

The close co-ordination of field and laboratory phases of the study has been a major concern in evaluating spectrographic results. Assembly of detailed information in the paper, and interpretation of data obtained, have been the responsibility of Mr. Graf. Incorporation of this study into the general pattern of the Santa Rita program of mineral investigation, and correlation with other studies, has been undertaken by Mr. Kerr.

Acknowledgments

This study received the support and wholehearted cooperation of the Kennecott Copper Corporation, the Empire Zinc Company, and the American Smelting and Refining Company. Their further courtesy in agreeing to make the report available for publication is gratefully acknowledged.

The Santa Rita program was the outgrowth of conversations with Mr. Anton Gray, chief geologist of the Kennecott Copper Corporation, who has kept in touch with the work from the start. The cooperation of Mr. W. M. Loerpabel, manager of the American Smelting and Refining Company, in making the facilities of the company available are greatly appreciated. Valuable advice has been given by Mr. William Callahan, manager of the New Jersey Zinc Exploration Company.

Mr. A. D. Hoagland and Mr. John Faick, geologists for the Empire Zinc Company, were most helpful in providing a clear picture of the geology of the Hanover area. Work at the Oswaldo mine was greatly aided by the suggestions of Mr. K. A. Strand, geologist of the Kennecott Copper Corporation for that property. Mr. George Ordonez, geologist with the exploration department of that company, supplied information on the Georgetown area and spent time with Mr. Graf in the field there. Mr. James Pollock and Mr. Thomas Mitcham, local geologists for the American Smelting and Refining Company, kindly discussed the geology of the Groundhog mine and the Tertiary area.

Dr. T. Ivan Taylor of the Department of Chemistry, Columbia University, made available the spectrographic equipment of that department, and made many helpful suggestions. The courtesy of the New Jersy Zinc Company in allowing Mr. Graf to spend several days at their research laboratory in Palmerton, Pennsylvania is deeply appreciated. Dr. M. L. Fuller and Mr. G. W. Standen discussed many

points of spectrographic technique during that visit. The suggestions received from Dr. L. H. Ahrens, of the Massachussett Institute of Technology, and Dr. L. W. Strock, of the Saratoga Springs Commission, Saratoga Springs, New York, were very helpful. Dr. J. L. Kulp and Mr. H. D. Holland of the Department of Geology, Columbia University, have reviewed portions of the manuscript.

The authors are especially grateful to Dr. T. S. Lovering for his careful reading and criticism of the paper.

Study of Trace Element Distribution

Procedures

Hydrothermal deposition of traces may result from decrease of temperature and pressure, direct chemical reaction between traces and wall rock, and neutralization of solutions by reaction with wall rock, resulting in precipitation of traces. Accepted concepts suggest that substitution in the lattices of wall-rock minerals should be greatest in deep-seated magmatic segregations, pegmatites, and hypothermal deposits where heat and pressure are high, and in altered rocks containing layer lattice minerals with high absorption and base-exchange capacities.

Most ore bodies are localized along linear or planar structural elements, such as faults, unconformities, joint intersections, and intrusive margins. Similarly, the "halo" of trace elements around an ore body may vary with the distribution of structural elements. However, where disseminated ore bodies occur in masses of highly altered or minutely fractured rock traces may diminish gradually in all directions. Even here, major structural elements are ordinarily responsible for localizing alteration and fracturing.

Traces deposited at high temperatures and under a heavy overburden may be expected to form a uniform aureole of limited extent, while traces in an epithermal environment should be widely distributed along fractures with minimum penetration of wall rock. The trace content of a specimen is therefore related not only to distance from the ore body and intensity of chemical reaction, but also to the kind and amount of fracturing. Increased fracturing away from an ore body could counteract a decreasing gradient predicted solely by chemical principles.

Wall-rock types contain trace element suites deposited when formed. These traces plus those introduced by ground water and minor periods of hydrothermal activity are to be distinguished from trace element suites associated with mineralization. An empirical approach to the problem of determining significance of distribution may be that of comparing spectrographic analyses of many specimens of the same rock type, at varying distances from ore, and considering the (low) metal content in most specimens as unrelated to mineralization. The *variation* of metal content in country-rock specimens unrelated to mineralization is also to be considered in determining a proper base value. A more fundamental but not necessarily more accurate approach is to determine the trace-contribution of separate non-mineralization events at places unaffected by mineralization, and to deduct these values from total trace content near ore bodies. A danger here is that, in securing specimens unaffected by mineralization, one may enter areas where conditions in the magma or sedimentational environment producing the original rock differed from those near the ore bodies. More important, this approach implies detailed knowledge of intrusive and hydrothermal history, often not available.

Two factors minimize these complexities: (1) Some elements occur almost exclusively in late residual concentrates of igneous activity. Such are much less abundant in igneous rocks and occur sparingly in sediments through erosion of pre-existing ore deposits, chemical precipitation, and concentration by organisms. The mineralization contribution of these metals should extend a considerable distance from an ore body before it becomes confused with pre-mineralization contents. (2) Characteristic gradients of concentration away from the ore body, rather than the mere presence of traces, are criticial. True, such gradients may also occur at intrusive margins because of assimilation or deuteric action. Likewise, original trace gradients in sedimentary rocks resulting from local concentration of detrital minerals and trace-rich organic remains, or from locally increased chemical precipitation, may be overlooked without petrographic study and a knowledge of

local paleogeography and paleoecology. In practice, these difficulties may be reduced by: (a) sampling near subsidiary ore bodies not coincident with intrusive contacts; (b) examining stratigraphic units having most nearly constant chemical composition, containing but few fossil types which are, however, abundantly distributed throughout the rock; (c) sampling a constant stratigraphic horizon, thus avoiding the cutting of time lines, unless many series of samples are to be collected.

Choice of Tracer Elements

A spectrographic analysis records the presence and approximate abundance of most metallic elements. This spread invites an empirical approach to trace element prospecting, in which a distribution-plot for each element is studied with regard to the location of known ore bodies. Many elements are unrelated to mineralization, however, and an empirical program becomes both tedious and confusing. The following criteria have been useful in predicting the utility of elements as tracers:

1. The tracer element should be restricted in geologic occurrence. Ca and Si are unlikely choices because calcite and quartz are prominent primary constituents of wall rocks, and may be introduced by ground water or hot spring activity as well as by hydrothermal solutions. However, Mg in a calcitic limestone may be used for prospecting where dolomitization is closely associated with ore. The ratio of the amount of the element introduced to that already present is critical.

2. Tracer elements need not be the main metals of the ore body, but must be associated. Be would probably not be useful in seeking galena-sphalerite ore bodies, for it is not normally deposited by sulfide-rich hydrothermal solutions. Zn and Pb would be satisfactory, as would Ag and Cd.

3. The tracer element is more likely to extend a useful distance from the ore body if it is one of the main ore metals. The limit of sensitivity of the spectrographic method without chemical enrichment of a sample is 0.0001% for most favored elements. An element comprising 0.1% of the hydrothermal mineral assemblage will not be detected when that assemblage constitutes less than 0.1% of a specimen. A minor metal may, of course, be deposited in greatest quantity away from the ore body, because of lower temperatures.

4. The tracer element must be present in enough samples to demonstrate a gradient or zoning. The absence of an element from a zone may be significant if it exists in other zones for comparison.

5. The tracer elements must have been deposited contemporaneously with the ore minerals. If late-ore or post-ore barite is deposited over a larger area than were ore minerals, a gradient of Ba concentration may point to a channel which conveyed Ba-rich solutions, but not ore.

6. The tracer element compounds must resist dilute acids, so that ground water will neither dissolve nor enrich them. Sokoloff (1948) states that compounds of Mo, W, Sn, and Pb are desirable for study because of generally low solubility. Zn compounds are among the most easily leached.

7. Sensitivity of the tracer element to the analytical method is important. The following, modified after Ryde and Jenkins (1946), gives the minimum percentage of an element necessary for detecting its strongest line using the the emission spectrograph.

0.0001%–0.001%	Ag, Co, Cr, Cu, In, Li, Mg, Na, Ni, Os, Pd, Pt, Rh, Ru
0.001%–0.01%	Al, Au, Ba, Be, Ca, Fe, Ga, Ge, Hg, Ir, Mn, Mo, Pb, Sc, Sn, Sr, Ti, V
0.01%–0.1%	B, Bi, Cd, La, Sb, Si, Tl, Y, Zn, Zr
0.1%–1.0%	As, Cs, Cb, K, P, Rb, Ta, W

Other things being equal, it should be easier to detect a gradient of Ag concentration than one of W, for the former is 1000 times more sensitive than W.

Sampling methods

Hand specimens were taken at increasing intervals (i.e., 0 feet, 0.5 foot, 1 foot, 2 feet, 5 feet, 10 feet, 25 feet, 50 feet . . . total distances from start of traverse) from surface and underground ore bodies and fault zones. Sampling intervals vary in different traverses because samples were taken only from rock defi-

finitely in place. In long intervals between major features, it was impossible to test for trace element gradients bordering every minor plane of movement, and specimens were taken at arbitrary paced intervals, with notes on fracturing and mineralization at collecting points. The pattern of drill core sampling was limited by the location of previously drilled holes.

In order to secure representative portions of field specimens without contamination by grinding, the slabs were broken with a hammer on a steel plate to obtain cross-sections through the specimen. About 15 grams of chips were then spalled from the surfaces, broken down in a large agate mortar, quartered, and reground.

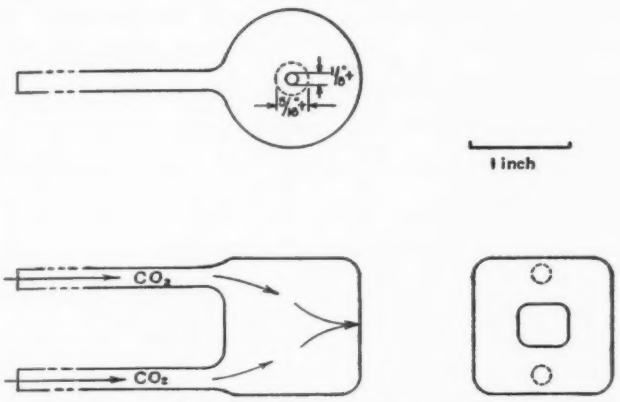


FIGURE 1.—FUSED QUARTZ HOOD USED TO INTRODUCE CO_2 ATMOSPHERE
The inlet tubes are 4½ inches long, so that rubber tubing may be attached to the outer end without burning.

SPECTROGRAPHIC PROCEDURES

Baird Associates' non-recording densitometer and their 3-meter grating spectrograph, with a dispersion of 5.6 Å/mm. in the first order, were used. A fused quartz hood (Fig. 1) was placed around the electrodes and arcing carried out in a slowly-moving carbon-dioxide stream, to repress cyanogen emission.

Crinoidal limestone samples, prepared by grinding together 30 mg. limestone and 30 mg. charred-sucrose carbon, were arced for 12 seconds at 12 amperes D C in center-pole graphite electrodes (Fig. 2) (Standen, 1944). Ca, an essentially constant major constituent in the Crinoidal, served as an internal standard.

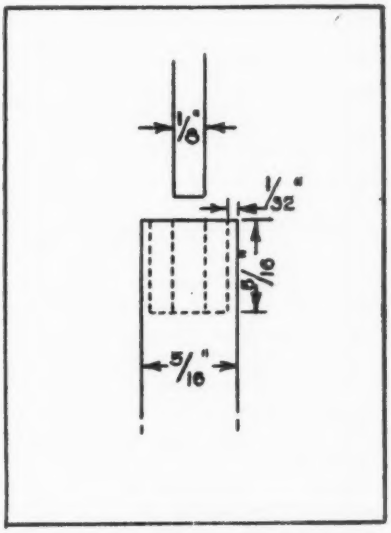


FIGURE 2.—THE CENTER-POLE ELECTRODE

A few shaly limestones and garnet specimens were analyzed in this way, but with indium chloride added as an internal standard. This procedure gives determinations for both refractory and non-refractory elements, during the initial period of arcing when carbon represses selective volatilization.

Finely-grounded 25 mg. samples of Tertiary clays and latites were arced at 10 amperes D C

for 45 seconds in central crater graphite electrodes, with indium chloride added as internal standard. This arcing period includes the volatilization peaks of metals like Pb and Zn, but does not give reliable determinations for refractory elements.

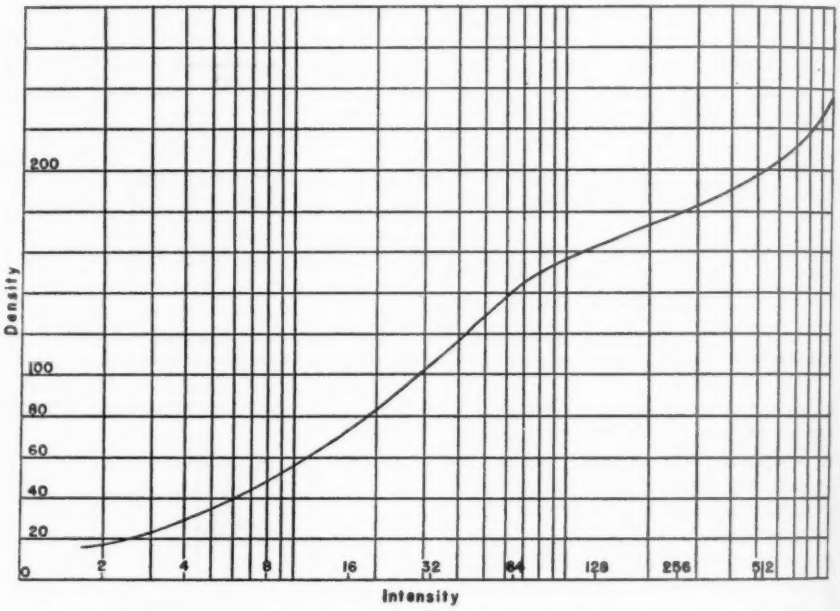


FIGURE 3.—DENSITY/INTENSITY CURVE FOR KODAK NUMBER 33 SPECTROGRAPHIC PLATES

Determinations of trace quantities in artificially-enriched C. P. $CaCO_3$ are reproducible to within 15 per cent of the arithmetic average.

"Intensity values" on graphs in this paper are relative numbers obtained by converting densitometer readings to intensities, and then correcting for background and for internal standard variation. Absolute concentrations on the graphs were obtained by arcing known quantitative mixtures and calculating their intensity values. The absolute values are necessarily approximate because, unless samples are completely consumed, the arcing behavior of a physical mixture cannot duplicate that of a natural specimen in which part of the metals are present as separate compounds and part within lattices of rock-forming minerals. Intensity-density relationships on films are shown in Figure 3.

LOCATION AND GENERAL GEOLOGY

The localities studied (Fig. 4) lie from 10 to 20 miles east of Silver City, New Mexico, in the Santa Rita Quadrangle. A belt of zinc mineralization extends from the Empire Zinc Company mine at Hanover, through the Oswaldo mine, to the Groundhog mine. The Georgetown and Lone Mountain districts are respectively northeast and southwest of this belt. An area of Tertiary outcrops lies to the south and east.

A few widely-spaced faults with throws measured in thousands of feet trend northwesterly, roughly parallel to the broad folding of the rocks (Paige, 1916). More closely spaced, northeasterly trending faults are generally much shorter, with throws only a fraction of those in the other group.

SEDIMENTARY ROCKS: The stratigraphic section in Figure 5 is complete except for a succession of Tertiary gravels, tuffs, and quartz latite flows. Spectrographic samples discussed in this paper come from the Tertiary, the Lake Valley formation, and the Oswaldo formation.

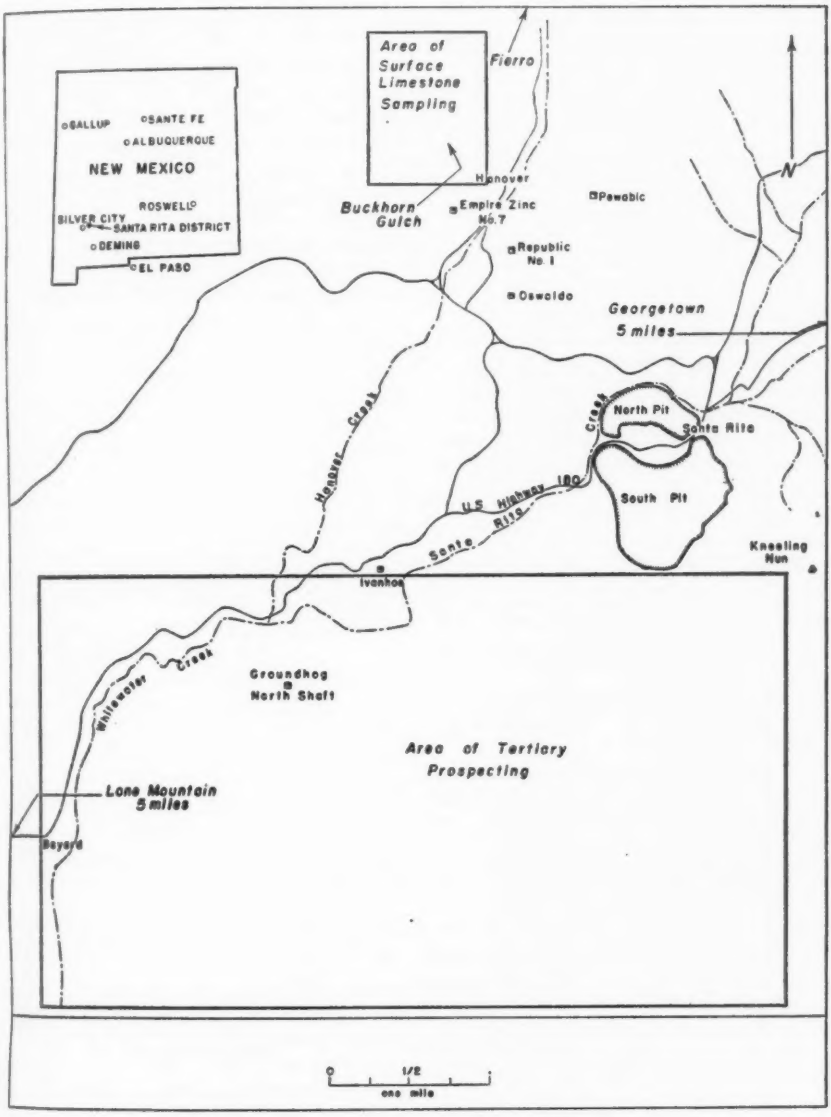


FIGURE 4.—INDEX MAP, SANTA RITA DISTRICT

The Lower Blue limestone, basal member of the Lake Valley formation, is fine-grained, dark gray, and medium-bedded, with many shaly horizons. Locally it contains rather pure, white limestone near the base. The overlying Crinoidal ("Hanover") limestone is white and coarse-grained, with white chert near top and bottom. It is normally composed largely of crinoid stems.

The Pennsylvanian Oswaldo formation has at its base the 20-foot arenaceous, light yellowish-brown Parting shale. The overlying Middle Blue limestone member is medium gray, with rather common shaly layers and

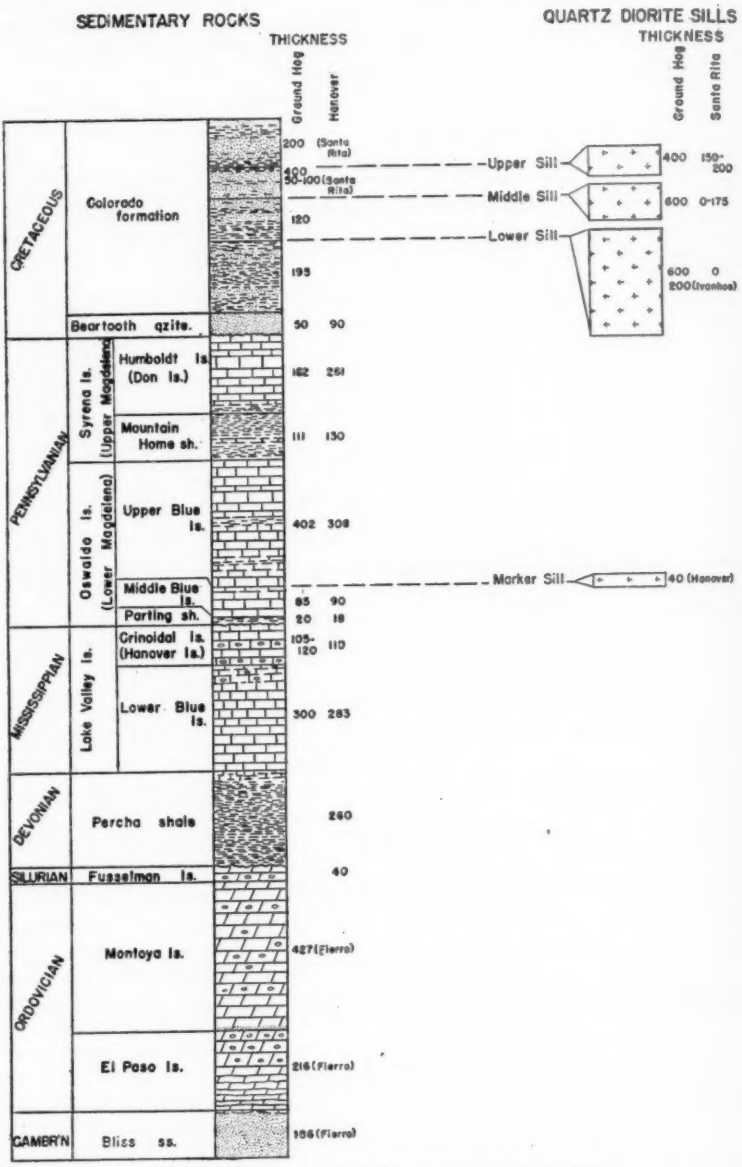


FIGURE 5.—STRATIGRAPHIC SECTION, SANTA RITA AREA

black chert lenses. The Upper Blue limestone, similar to the Middle Blue lithologically, is separated from it by the 40-foot cretaceous quartz diorite "Marker" sill.

IGNEOUS ROCKS: The Cretaceous sills (Fig. 5) were injected prior to a regional synclinal warping (Lasky, 1936). The Bayard-Groundhog area lies on the eastern flank and near the

trough of this syncline. Intrusion of granodiorite stocks at Santa Rita and Hanover followed folding. Injection of late Cretaceous to early Tertiary dikes completed the main intrusive history. Tertiary sedimentation, extrusive activity, and intrusion of quartz latite dikes ensued.

Wright and Kerr (1947) recognize three dike types throughout the Santa Rita area, which correlate and simplify geological nomenclature throughout the district. They are Cretaceous hornblende (diorite) porphyry, later Cretaceous quartz (diorite) porphyry, and Tertiary quartz latite. Absence of epidote in the quartz latite dikes dates them as post-ore, for the main period of epiditization coincides closely with ore deposition. The epidotized hornblende-porphyry dikes are definitely pre-ore at all localities. The quartz-porphyry dikes, which cut the hornblende porphyries, were intruded over a time range essentially contemporaneous with ore deposition. They contain sphalerite at one place in the Oswaldo mine, but no ore and very little epidote at Hanover. At the Groundhog some contacts against silicates are chilled, while others are epidotized.

The two Cretaceous dike types at Hanover generally follow the prominent northeast-trending fault zones, while many quartz latite dikes cut at right angles. Extreme variation in strike of the later dikes, and their marked change in thickness along strike, suggest a changed regional stress pattern after ore deposition. The same cross-cutting of quartz-latite dikes with regard to quartz porphyry dikes has been noted at Santa Rita (Patterson and Kerr, 1947); and another trends east-west across the Tertiary plateau east of the Groundhog mine.

Tertiary Area

General Statement

The Tertiary quartz latite flows, tuffs, sands, and gravels immediately east of the Groundhog mine dip moderately westward, and several hundred feet farther east they assume a regional dip of 0°–12° southeastward. West of the Groundhog fault (Fig. 6), San Jose Mountain and lake gravels in Bayard Canyon constitute a Tertiary remnant in an area of Cretaceous Colorado formation and interbedded sills. Lasky (1946) and Kent (1948) give more complete stratigraphic and structural descriptions.

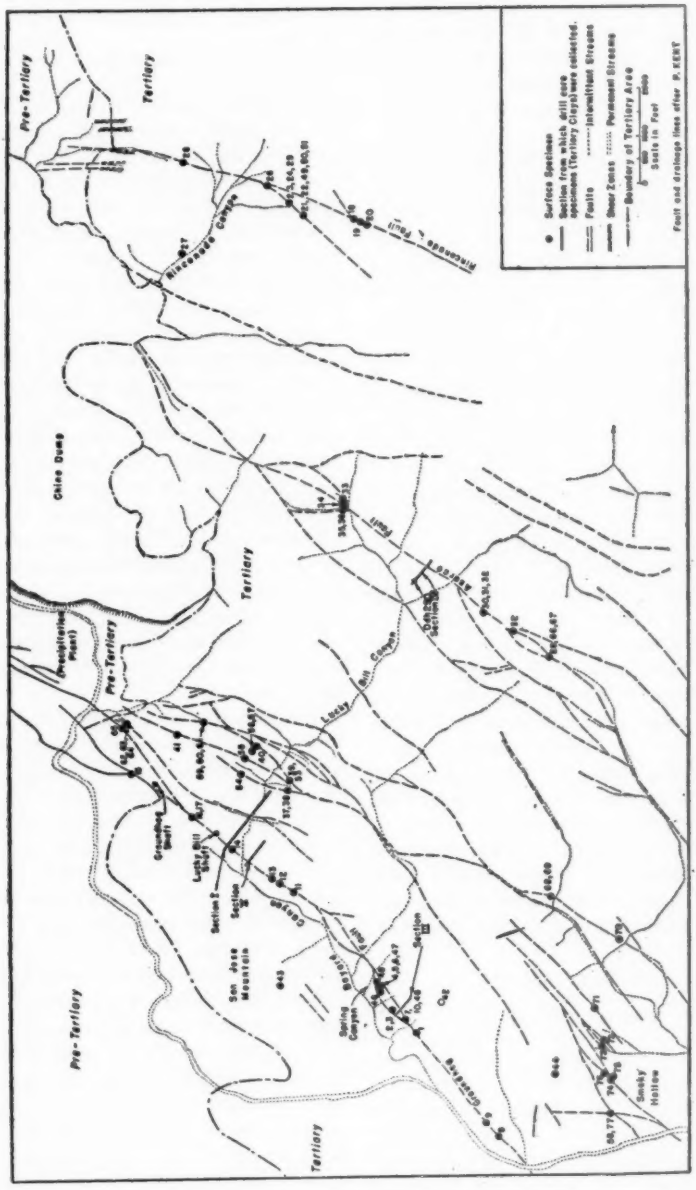
Prospecting in the Tertiary

The Tertiary section covers the dikes, sills, and pre-Tertiary sedimentary rocks that are hosts for mineralization at the Groundhog mine. Prospecting thus far for ore bodies beneath the Tertiary has been based upon two observations: (1) Faults in the Tertiary trend parallel with those in nearby Cretaceous outcrops. (2) The Groundhog fault, which cuts both Tertiary and earlier rocks, has a 250-foot vertical component of Tertiary movement, a significant fraction of the total 1500 feet of vertical movement noted by geologists of the American Smelting and Refining Company. It is inferred, therefore, that renewed movement after Tertiary vulcanism, along existing faults in Cretaceous and older rocks, resulted in displacements of the Tertiary. Conversely, major present-day Tertiary faults may be extensions of faults in the Cretaceous below. The latter have been shown capable of channeling ore-bearing solutions. The problem is complicated, however, by instances encountered in drilling where pre-Tertiary faults stop under the Tertiary.

The thickness of Tertiary cover makes it costly to drill to the Cretaceous rocks, which presumably might contain ore in brecciated dike margins or jasper veins, even in valleys where part of the Tertiary has been eroded. Possible replacement bodies in limestone lie even deeper, and it might not be feasible to test for them along a fault unless there was strong evidence of mineralization in Cretaceous rocks. Therefore, some special criterion is needed to determine factors favoring mineralization along one fault channel versus another, and along different parts of the same channel.

In mapping the Tertiary area, Kent (1948) aimed to delineate faults, measure their displacement, and study replacement effects. That program revealed no visible sulfide minerals along the faults. Further petrographic work by Kent should determine types of alteration.

A preliminary trace element study, supplementing field and petrographic studies, was undertaken to indicate spots along fault planes


Pre-Tertiary
Tertiary
China Draw
Precipitation Point
Grayback Draw
Ladder Ridge
Section 2
Section 1
Section 3
Section 4
Section 5
San Jose Mountain
Spring Canyon
Section 6
Section 7
Section 8
Section 9
Section 10
Section 11
Section 12
Section 13
Section 14
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Section 16
Section 17
Section 18
Section 19
Section 20
Section 21
Section 22
Section 23
Section 24
Section 25
Section

meriting more intensive trace element work and perhaps drilling. This program was divided into: (1) collection of specimens of pink and white Tertiary clays from drill cores, and (2) collection of specimens of tuffs and latites cropping out near prominent faults, especially where the latite is bleached. In spite of the lack of visible sulfides in the Tertiary area, and Lasky's statement (1936) that Tertiary mineralization was a minor aftermath of the main period, trace element concentrations might be detected because: (1) Later hydrothermal solutions rising through a channel coated with ore minerals should dissolve some and redeposit them at levels of lower temperatures. (2) Ground water under artesian head might transport traces upward from an ore body and redeposit them. (3) The Tertiary solutions may have carried a distinctive group of trace elements themselves. In the third case, the quantity of traces found along fault zones would indicate volume of Tertiary solution flow, even though the solutions were not ore depositing. Although it has not been possible to carry out field work to evaluate the second possibility, the correlation of trace highs with known ore bodies described later is most readily explained by hydrothermal transport of traces followed by some ground-water redistribution.

Bleached Zones

At points along faults, the latite is bleached and disintegrating into a coarse sand. Petrographic examination verifies the freshness of the feldspars and the absence of significant alteration minerals except powdery white films of secondary quartz. Hand specimens from the Groundhog fault, just east of Bayard, contain pinkish quartz veinlets extending back from the strongly slickenslided surface of the latite. The bleaching thus results from addition of films of whitish quartz, and disintegration of a thin outer layer of the rock. It is somewhat uncertain whether these effects are caused by Tertiary hydrothermal solutions, by secondary solutions derived from tuffs in the Tertiary section, or merely by weathering of badly broken zones.

Pink Clay Horizons

Tuff zones partially or completely altered to pink and white montmorillonite occur in the lower part of the Tertiary section, interbedded with sands and gravels. There is no significant difference in optical properties of pink and white clays. The pink color is caused by Mn, for spectrographic analysis indicates high (spectrographic) quantities of Mn comparable to those found for white clays containing dendritic black spots of Mn oxides.

Tuffs over much of the Tertiary area are partly altered to montmorillonite, mottled in white and shades of pink. In drill core available, however, it seems that crumbly deep-pink clays and white clay rock containing black Mn oxide spots are localized along faults. This suggests that the Mn oxides were not deposited with the Tertiary sediments, but were secondarily introduced by circulating water.

Approaching an altered tuff zone in core, black Mn oxide specks are first noted, then pink haloes of increasing size around the oxide centers appear, until presently the clay is deep pink and the centers have disappeared. Likewise, a gradual transition is noted from tuff, to clayey rock with residual tuffaceous structure and quartz and biotite grains, to a virtually pure clay rock. Thus, three solution effects are observed: (1) conversion of tuff to clay; (2) deposition of Mn oxides; (3) dispersion of Mn to form pink clays. The first two may have been contemporaneous. The third could result simply from ground-water activity.

Spectrographic Data, Tertiary Surface Specimens

Study was concentrated upon slickenslided dark-brown films, secondary quartz, and gouge along actual fault channels. Several composite samples from within zones of disseminated movement were examined, as were a few specimen sets from traverses perpendicular to faults. Most of the latter specimens proved most useful as standards for trace content of the original rocks, for trace addition from solution was spatially restricted except in tuffs. (See Fig. 7 for summary.)

1. Outstanding Zn maxima are in specimens 66 and 67, gouge and breccia from a prospect

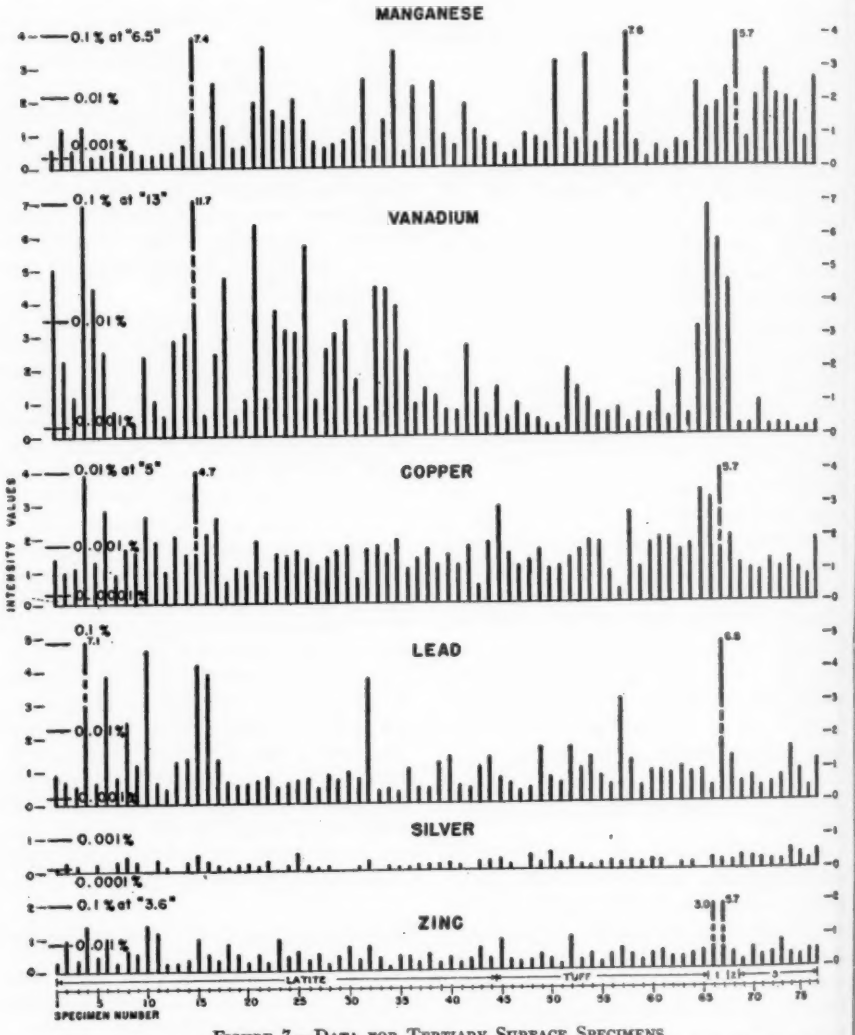

FIGURE 7.—DATA FOR TERTIARY SURFACE SPECIMENS

Rock type "1", gouge and breccia of uncertain composition; "2", greensand; "3", silicified rhyolite. Interfering lines of other elements masked prominent Ag lines in specimens for which no Ag is reported. Relative intensity values for different elements are not directly comparable.

on the Asarco fault. The low metal content of specimen 29, a disintegrated latite only 3 feet from the gouge specimens, indicates the limited distance over which trace addition took place. Rocks at this prospect are badly broken and iron-stained, but specimen 21 from a fault zone in Rinconada Canyon is similarly stained and fractured, and has a low metal content.

The Zn content of several bleached latites from the Groundhog fault (Specimens 4 and 10, for example) is slightly higher than average, but the difference is within the limit of experimental error.

2. There are no significant Ag highs.

3. Pb highs come from gouge and breccia specimens 66 and 67, from a sheared latite (57)

along the Asarco fault near numbers 66 and 67 and from bleached latites along the Groundhog fault (4, 6, 10, 8, 15, 16). The isolated Pb high for specimen 57, an otherwise low metal-bearing tuff 45 feet from a minor fault, cannot be regarded by itself as significant.

The first three of the Groundhog bleached-latite highs are from an area where drilling has located sphalerite in underlying limestone, the last two from points above existing mine workings. Even in these favored areas, Pb distribution is erratic, for some latites (3, 7, 14) and all tuffs (45, 48, for example) give low Pb values. Some doubt may exist about validity of highs in specimens 15 and 16, which were collected from outcrops near mine dumps. The occurrence of Pb anomalies correlatable with known ore bodies, in some latite specimens that are fresh and massive except for a thin bleached layer, indicates that Pb must be present in surface films. This contrasts with Zn, which was either never deposited in Groundhog latites or was leached, apparently because of the absence of clayey gouge like that making up specimens 66 and 67.

Latites (33, 34, 35, 36) from a bleached zone (Section IV, Fig. 6) have Pb values definitely lower than those of the Groundhog bleached latites. There may not be comparable mineralization beneath the two points. Another factor is the difference in vertical distance which traces would have to move from ore bodies, for the Groundhog fault moved pre-Tertiary ore bodies west of it nearly into contact with Tertiary rocks east of it.

4. Cu highs in latites correlate well with Pb highs. Those in tuffs are not geographically localized and do not correlate with other metal highs; they are of no apparent value as tracers.

5. V/values are extremely variable, but of a different magnitude for tuffs than for latites. V highs coincide with some Pb peaks for bleached latites (4, 6, 8, 10, 15, 16) and for the two Zn highs (66 and 67). However, the failure of so many V highs to correlate with Pb and Zn peaks, whose close relation to ore bodies has been shown, suggests that weathering or alteration of Tertiary specimens is critical in releasing V from rock-forming minerals so that it is more easily volatilized during spectrographic analysis.

6. Mn highs show a negative correlation for Pb and Cu highs. There are only two undoubted Zn maxima, not enough for comparison. Although the genetic relation between Pb – Zn and Mn is discussed further under the data for Tertiary clay specimens, it seems obvious that Pb and Zn (especially Pb) are more closely related to known ore deposits than Mn, and should be the basis for further spectrographic prospecting.

7. The silicified rhyolite unit (Kent, 1948) (Specimens 69–77), virtually a chalcedony rock in places, has received the greatest addition of material of any Tertiary rock. However, the low trace element content of these specimens, except for Mn, indicates that silicification is not significant in prospecting.

8. Tuffs fail to give useful trace concentrations, even though they are permeable and usually contain clay. Perhaps channeling of solutions into fault zones of limited volume in the latites, and resulting increased intensity of solution activity, created bleached zones having a greater trace content than the tuffs.

Spectrographic Data, Tertiary Clays

Tertiary drill core samples were obtained: (1) near mines in the pre-Tertiary (Figs. 6, 8, 9); (2) near lesser, unexploited pre-Tertiary mineralization (Figs. 6, 10); (3) from drill hole 290 in the Tertiary (Figs. 6, 11). Spectrographic data are shown in Figure 12.

Specimens of high metal content have been regrouped (Table 1), in decreasing order for each metal. Some conclusions in the discussion following are based upon clay content, color, and texture of individual specimens, information too bulky to be reproduced.

1. Cu highs are not localized with regard to ore bodies. Their localization in clayey sandstones, rather than in pure clays, suggests that much of the Cu may be an original constituent of the sandstone.

2. V highs are concentrated in Sections I and II, but correlate poorly with regard to ore bodies in those sections. As with Tertiary surface specimens, it seems likely that V in rock-forming minerals has been extensively released in areas of greatest solution movement.

3. Mn highs occur throughout the entire horizontal extent of the sections studied, in relatively permeable, thin-bedded clayey tuff-

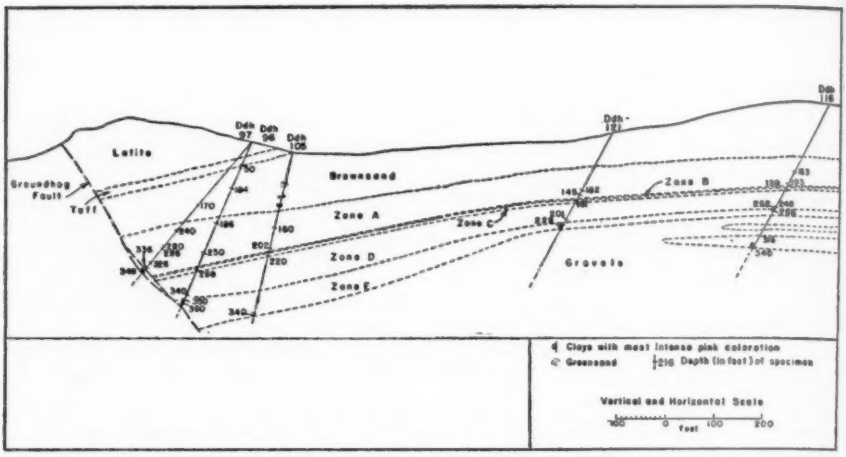

FIGURE 8.—TERTIARY CLAY SPECIMENS, SECTION I

Zone A. Thin-bedded sandstones, clayey sandstones, clays, and occasional unaltered tuffs.
Zone B. Bluish-gray silica fragments in white clay.
Zone C. Thin sandstone, tuff, and gravel-rich clay zones.
Zone D. Uniformly fine-grained, light gray-green sandstone, tuffaceous toward the base.
Zone E. Like Zone C.
The two lenses in drill hole 116 are thin-bedded clayey sandstones.

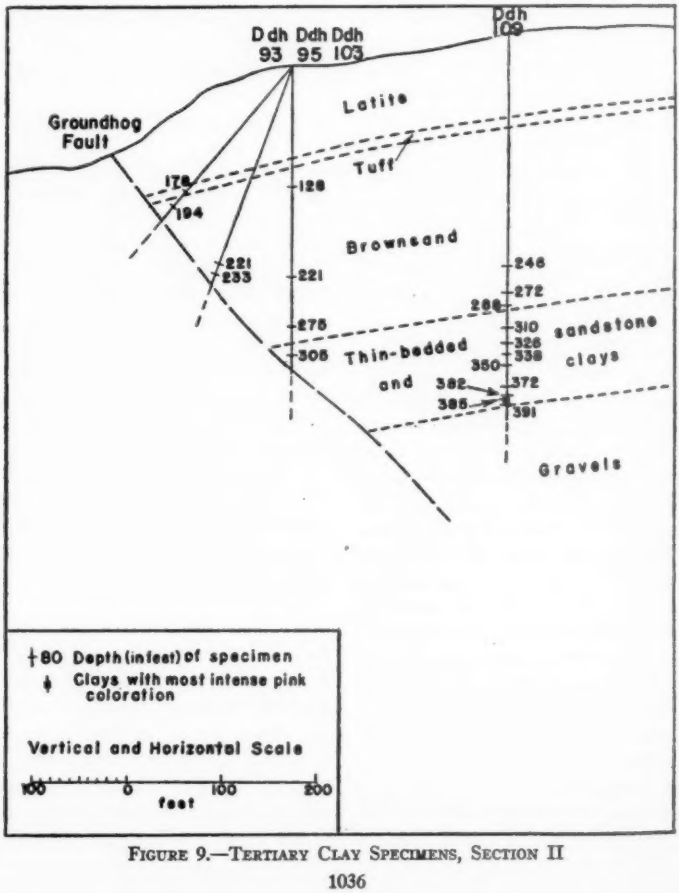

FIGURE 9.—TERTIARY CLAY SPECIMENS, SECTION II

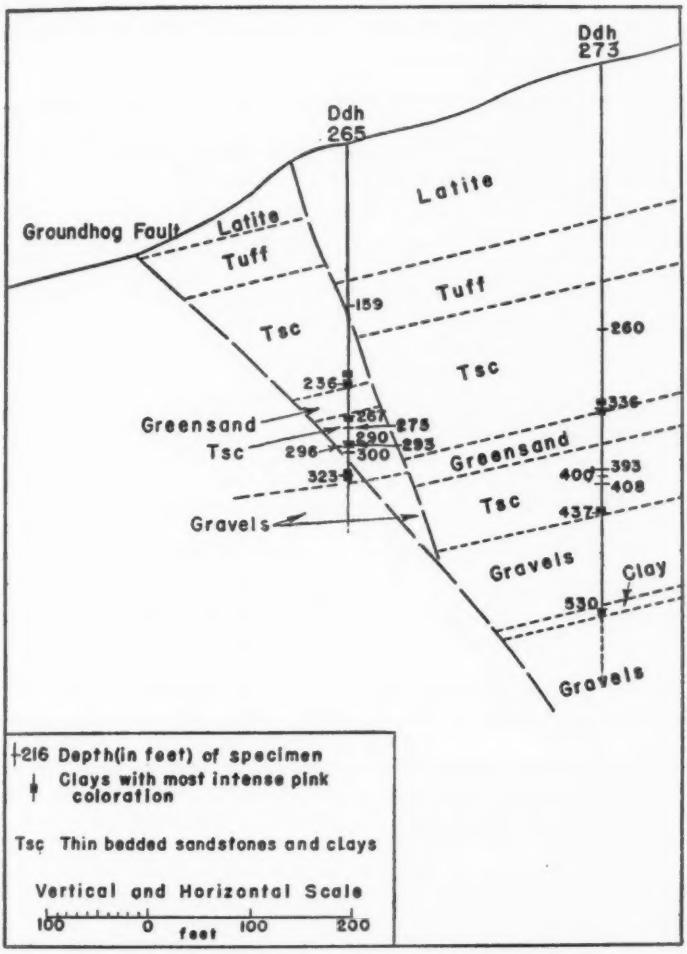

FIGURE 10.—TERTIARY CLAY SPECIMENS, SECTION III

sandstone alternations, rather than in thick sandstones. Mn highs for pink clays, and porcelainous white ones containing Mn oxides, are comparable.

4. Ag highs occur within 20 feet of the Groundhog fault in an area above producing mines, as a zone which cuts across Tertiary beds. These specimens are generally high in Pb and Zn.

5. The most prominent Zn highs are from Sections I and II, lesser ones from Section III, and none from Section IV. Most are within 60 feet of the Groundhog fault, but Specimens 105–340, 105–160, and 109–288 are exceptions. The two highs toward the east end of Section I may have been contributed by solutions rising along fractures lying farther east.

Extreme Zn highs are found in porcelainous white clays, lesser peaks in deep pink clays. Correlation with other metal highs is poor, probably because of solubility and consequent redistribution of Zn compounds.

6. Pb highs are absent from Sections III and IV. In the other two sections, they occur

in a zone extending about 40 feet from the Groundhog fault, with one exception (96–340). Independent of the degree of pink coloration of the clay, they are found in all stratigraphic horizons, and correlate best with Ag, Zn, and Mn highs.

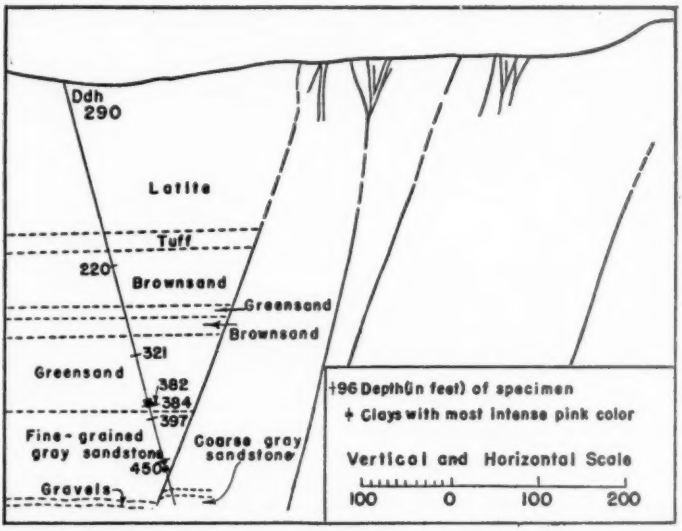

FIGURE 11.—TERTIARY CLAY SPECIMENS, SECTION IV

Ag, Zn, and Pb were clearly introduced into permeable Tertiary horizons by solutions moving up the mineralized Groundhog fault. In Sections I and II, Ag was deposited near the fault, Pb through a wider zone, and Zn even farther. There is only a restricted zone of highs in Section III, and no highs at all in Section IV. It is not clear why the Ag zone should have been more restricted than that of Pb.

Localization of the parent Mn oxide specks, from which pink color spreads, to porcelainous white clays suggests that Pb-Zn-Ag and Mn may have been contemporaneously introduced. Mn could have been derived from Mn-bearing silicates associated with replacement ore bodies in limestones below the Tertiary.

Occurrence of outstanding Zn highs in porcelainous white clays containing black Mn oxide specks indicates ground water has not penetrated these specimens to disperse either metal. In zones of crumbly, porous, deep-pink clay, there is some indication that highest Mn values occur with low Zn, and vice versa, which needs to be compared with experimental data on absorption of Zn and Mn by montmorillonite.

It should be emphasized that the "zones" of high metal content described contain a considerable percentage of rock low in these metals. Metal-laden solutions moved through irregular channels of higher porosity. The balance between rate of metal supply, rock surface in contact with solutions, temperature decline, solution velocity, etc., was such that the supply of a given metal was exhausted at a definite distance from the fault channel.

The negative correlation between Mn and Pb highs in latites must be related to the absence of fine-grained clay which could retard Mn removal, for these elements occur together in the clay samples. Likewise, surface tuff specimens were apparently collected in areas where solution activity was not intense enough to convert them fully to clay or to deposit traces comparable to those found in core near the Groundhog fault.

PRE-TERTIARY ORE BODIES

The Empire Zinc and Oswaldo Mines

The Empire Zinc Company mine lies along the southwestern margin of the Hanover Cre-

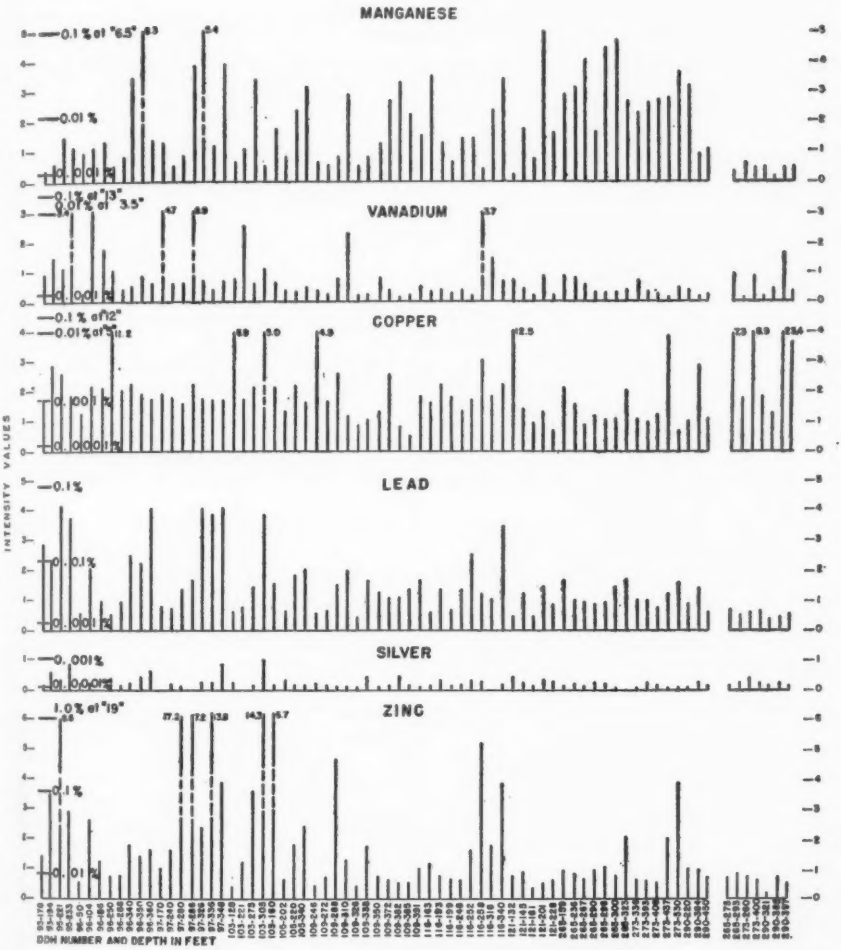

FIGURE 12.—DATA FOR TERTIARY CLAY SPECIMENS

Interfering lines of other elements masked prominent Ag lines in specimens for which no Ag is reported. Relative intensity values for different elements are not directly comparable.

taceous granodiorite porphyry stock. Deformation accompanying emplacement of the stock created a marginal syncline in the sedimentary section, succeeded by a nearly symmetrical anticline which merges into a homocline dipping southward about 15°. Zones of repeated movement, intruded by all three dike types, radiate from the southern end of the intrusive. The Kennecott Copper Company's Oswaldo mine lies entirely on the homocline, but some Empire Zinc workings extend into the marginal folds.

The Marker Sill at Hanover is cut by a hornblende-porphyry dike, and both dike and sill are cut by the Hanover intrusive (Schmitt, 1939). The three rocks were later pyrometasomatized during silication immediately preceding ore deposition. Post-ore dikes which cut all these rocks are not pyrometasomatized, but contain brecciated fragments of silicates and ore plucked from the walls by the dike magma.

The Crinoidal limestone is the main host for ore deposition, but commercial mineralization is also found in the shaly Blue limestones. Only

TABLE 1.—TRACE ELEMENT HIGHS IN TERTIARY CLAY SPECIMENS

Specimen number	Section & distance from fault in feet	Stratigraphic zone	Content of other metals
Zinc			
97–280	I 60	Zone A	.....
97–336	I 5	Zone A	H-Pb
103–305	II 10	Thin-bedded sandstones & clays	H-Ag, H-Pb, H-Cu
95–221	II 20	Brownsand	H-Pb, M-Cu
97–286	I 55	Zone A	H-V, H-Mn
105–160	I 290	Zone A	.....
116–258	I 1225	Zone E	M-Cu, H-V
109–288	II 255	Brownsand	M-Pb, M-V
97–348	I 5	Zone A	H-Ag, H-Pb M-Mn
116–340	I 1150	Ss. lens in gravels	H-Pb, M-Mn
273–530	III 100	Clay	M-Mn
103–275	II 35	Brownsand	M-Mn
93–194	II 20	Brownsand	H-Ag, M-Pb, M-Cu
96–104	I 225	Brownsand	H-V
105–340	I 120	Zone E	M-Mn
265–323	III 10	Thin-bedded ss. & clays	M-Mn
273–437	III 175	Thin-bedded ss. & clays	H-Cu, M-Mn
Copper			
290–382	IV 55	Greensand	M-V
121–132	I 935	Zone A	.....
96–250	I 130	Zone A	.....
273–260	II 230	Thin-bedded ss. and clays	.....
265–275	III 25	Thin-bedded ss. and clays	.....
103–128	II 140	Brownsand	.....
103–305	II 15	Thin-bedded ss. and clays	H-Zn H-Pb
109–246	II 295	Brownsand	
273–437	III 165	Thin-bedded ss. and clays	M-Mn, M-Zn
290–397	IV 45	Gray ss.	.....

Specimen number	Section & distance from fault in feet	Stratigraphic zone	Content of other metals
Lead			
97–348	I 5	Zone A	H-Zn, H-Ag, M-Mn
97–336	I 5	Zone A	H-Zn
97–326	I 30	Zone E	M-Zn, H-Mn
96–360	I 10	Zone E	H-Ag
95–233	II 10	Brownsand	M-Zn, H-V, H-Ag
95–221	II 20	Brownsand	H-Zn, M-Cu
103–305	II 10	Thin-bedded ss. & clays	H-Zn, H-Ag, H-Cu
116–340	I 1150	Ss. lens in gravels	H-Zn, M-Mn
93–178	II 40	Tuff	.....
93–194	II 20	Brownsand	M-Zn, M-Cu, H-Ag
96–104	I 225	Brownsand	M-Zn, H-V
96–340	I 30	Zone D	M-Mn
96–350	I 20	Zone E	H-Mn
116–252	I 1225	Zone E	.....
Silver			
103–305	II 15	Thin-bedded ss. and clays	H-Pb, H-Zn, H-Cu
97–348	I 5	Zone A	M-Zn, H-Pb, M-Mn
95–233	II 10	Brownsand	M-Zn, H-Pb, H-V
93–194	II 20	Brownsand	M-Zn, M-Pb M-Cu
96–360	I 10	Zone E	H-Pb

"H", metal content is high compared to value for all specimens; "M", moderate; no notation is made if content is low.

TABLE 1.—(*Continued*)

Specimen number	Section & distance from fault in feet	Stratigraphic zone	Content of other metals
Manganese			
96–350	I 20	Zone E	M-Pb
97–326	I 30	Zone E	M-Zn
121–201	I 820	Zone D	.....
265–300	III 5	Thin-bedded ss. and clays	.....
265–296	III 10	Thin-bedded ss. and clays	.....
265–267	III 30	Thin-bedded ss. and clays	.....
97–286	I 55	Zone A	H-Zn, H-V
97–348	I 5	Zone A	H-Ag, H-Pb, H-Zn
273–530	III 100	Clay	H-Zn
116–163	I 1370	Zone A	.....
96–340	I 30	Zone D	M-Pb
103–275	II 35	Brownsand	H-Zn
109–382	II 170	Thin-bedded ss. and clays	.....
116–340	I 1150	Ss. lens in gravels	H-Pb, H-Zn
105–340	I 120	Zone E	M-Zn
290–220	IV 145	Brownsand	.....
265–236	III 25	Thin-bedded ss. and clays	.....
Vanadium			
97–286	I 55	Zone A	H-Zn, H-Mn
97–170	I 180	Brownsand	.....
116–258	I 1225	Zone E	H-Zn, M-Cu
95–233	II 10	Brownsand	M-Zn
96–104	I 255	Brownsand	M-Zn
103–221	II 80	Brownsand	.....
109–310	II 445	Thin-bedded ss. and clays	M-Mn
96–186	I 195	Zone A	.....
290–382	IV 55	Greensand	H-Cu
93–194	II 20	Brownsand	H-Ag M-Pb, M-Cu, M-Zn
116–316	I 1160	Ss. lens in gravels	M-Mn

minor mineralization occurs in the dolomitic El Paso, Montoya, and Fusselman limestones.

CONTACT METAMORPHISM AND METASOMATISM: Bordering the intrusive is a 200- to 800-foot belt (Schmitt, 1930) in which shales and non-magnesian limestones have been completely replaced by andradite garnet, hedenbergite (salite, according to Schmitt), and epidote, with magnetite, chalcopyrite, and sphalerite-ilvaite-pyrite-pyrrhotite occurring at points within the zone. The silicates formed first, magnetite deposition partially overlapped silicate formation and continued afterward, and the sulfides are youngest.

At the edge of the metasomatic belt there is an abrupt change to limestones, now recrystallized and decolorized, and shales largely converted to hornstone. This belt of metamorphism may extend 1200 feet farther (Schmitt, 1930).

Garnet and magnetite extend into the metamorphic belt along dike-fault zones, and there are also isolated centers of metasomatic replacement along fractures. Certain favorable

stratigraphic horizons, notably the calcitic part of the Lower Blue and the Crinoidal limestone just under the Parting shale, are frequently replaced by garnet and hedenbergite.

marked by a thin zone of pink to brown, Mn-Fe rich carbonate.

Blanket ore bodies under the Parting shale occur just outside the metasomatic zone, from

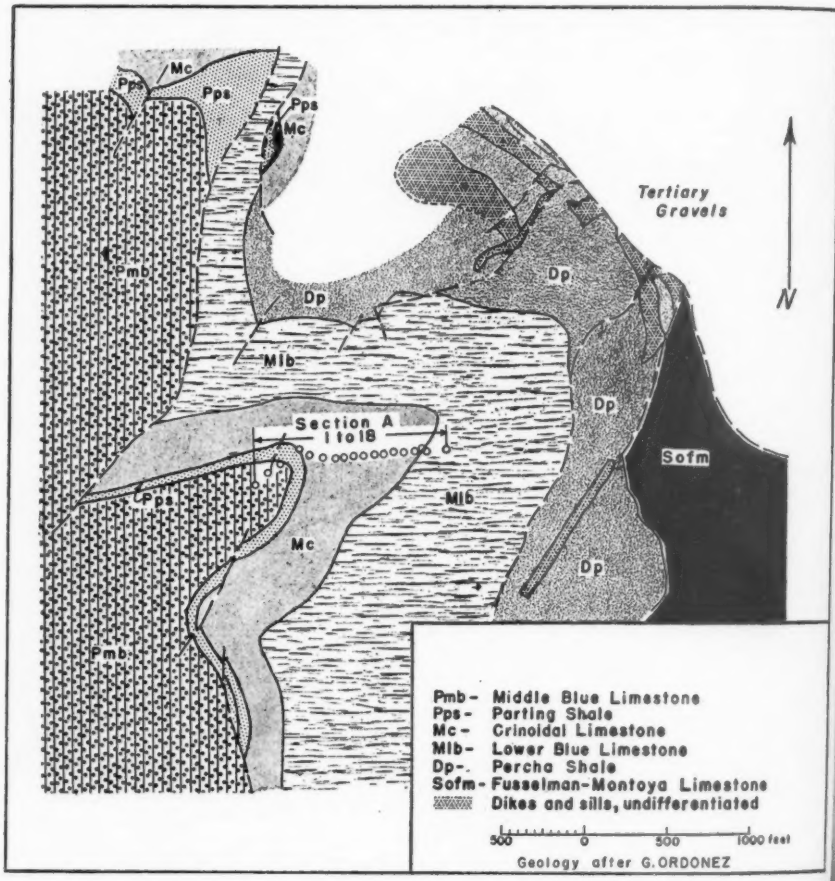

FIGURE 13.—LOCATION MAP FOR GEORGETOWN SURFACE SPECIMENS

ORE BODIES: Steeply dipping, sheet-like ore bodies lie on either side of dike-fault zones, widening appreciably under the Parting shale. Dike rock next to the fault usually contains epidote, but not sulfides. The limestone is replaced for 10 to 20 feet by garnet, hedenbergite, and minor sulfides, followed by a belt of high-grade ore containing some silicates. The contact of ore with recrystallized Crinoidal limestone, highly irregular in outline, is frequently

the fractured crest of the marginal anticli for several hundred feet down the west limb

The increase in galena in ore bodies farth from the Hanover intrusive (Schmitt, 193 indicates progressively decreasing temperatu of deposition.

Georgetown and Lone Mountain

Minor diggings at Lone Mountain, and number of rather extensive mines at Georg

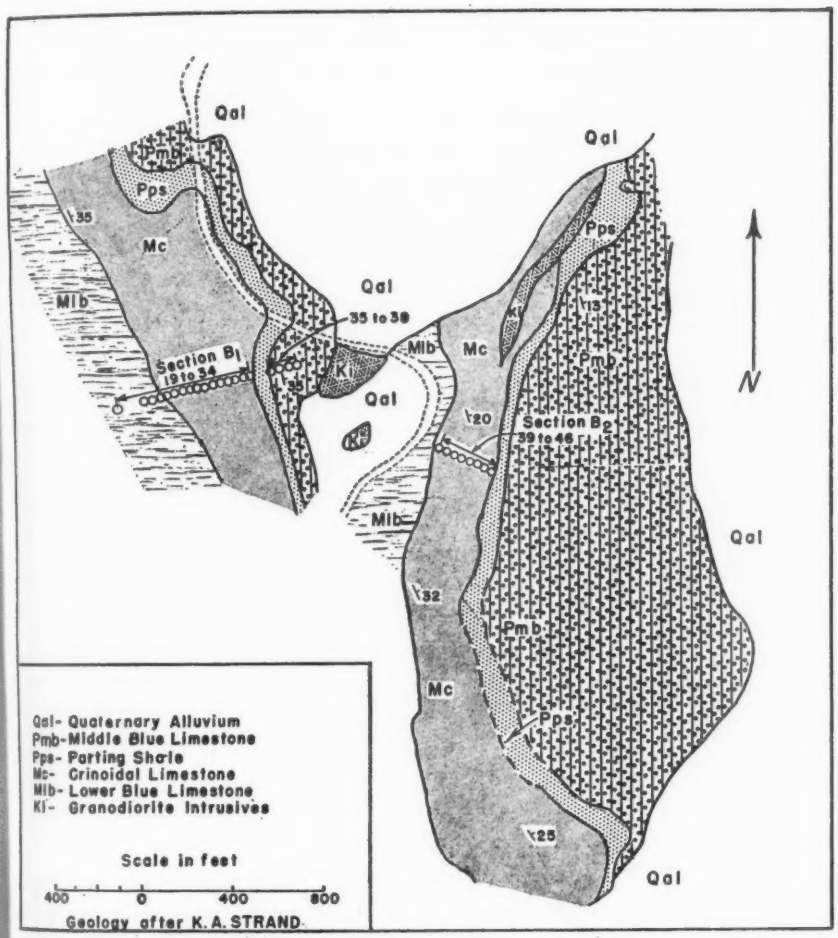

FIGURE 14.—LOCATION MAP FOR LONE MOUNTAIN SURFACE SPECIMENS

town, in the past exploited Pb-Ag-(V) ore bodies. The preference of this mineralization for the dolomitic Fusselman limestone, rather than the Crinoidal, has not been explained.

A few specimens were collected in the Crinoidal above these ore bodies at Georgetown (Fig. 13) and Lone Mountain (Fig. 14) to test whether a distinctive trace element contribution from the ore assemblage could be recognized.

The Groundhog Mine

The Groundhog mine of the American Smelting and Refining Company lies in the general vicinity of the Groundhog normal fault zone, which trends northeastward and dips about 50° to the east. The main fault is said to have a total displacement of about 3300 feet, with a vertical component between 1500 and 1600 feet. Recurrent movement along the fault has been outlined in considerable detail, for each movement formed a new set of subsidiary faults.

Three periods of hypogene mineralization have been recognized in the Bayard area (Lasky, 1936): (1) non-commercial quartz-pyrite earlier than any dikes, but later than and probably genetically related to intrusion of the

sills; (2) commercial veins of sphalerite, galena, and chalcopyrite emplaced during the main period of ore deposition, following intrusion of the quartz porphyry dikes; (3) minor quartz and pyrite later than the quartz latite dikes.

ORE BODIES: In the upper levels, ore occurs in altered dikes, sills, and Colorado formation adjacent to a vein of hematite-stained quartz-jasper breccia. The vein-type ore bodies result from continued reopening and filling of fractures, the fractures tending to follow dike walls (Lasky, 1936). Hypogene ore minerals are galena, chalcopyrite, and pyrite. Oxidation has developed rich near-surface bodies of cerussite and other secondary minerals of Pb, Zn, Mn, Cu, and V. These give way to a secondarily enriched chalcocite zone above the primary sulfide zone.

In replacement bodies below pre-ore silication is locally as intense as at Hanover, but less widespread. Hedenbergite is common, andradite garnet is restricted to the vicinity of main ore channels, and ilvaite is rare. Much of the ore has no silicates, but pyrite-quartz carbonate gangue instead. The shales, although often partly replaced by sericite and quartz near ore, are rarely converted to a white hornstone or partially replaced by garnet. Decolirization occurs only at a few points near ore, and even recrystallization of limestones is limited.

Spectrographic Data, Surface Limestone Specimens

These specimens (Figs. 13–17) represent a single traverse across the Crinoidal at Georgetown, two at Lone Mountain, successive traverses in Buckhorn Gulch going northward from the Empire Zinc open pits, and a suite collected near a magnetite mass in the gulch. Recrystallization at Hanover and Lone Mountain increased permeability slightly and partially lessened control of solution flow by fractures. At Georgetown recrystallization was minor, the rock is in places dense with only a few crinoid stems, and the Lower Blue-Crinoidal contact is gradational over many feet.

1. The decrease in Al content in many traverses, proceeding upward from the Lower Blue-Crinoidal contact, and the corresponding increase in Mn, are believed to be original sedimentary features. Excellent Mn and Al gradients at Lone Mountain, where there is no garnet or hedenbergite, rule out an origin during pre-ore silication. It is unlikely that the gradients stem from the immediate period of ore deposition, for other elements directly related to ore deposition (Pb and Zn) were not able to penetrate the Crinoidal any great distance. Furthermore, Mn and Al in ore bodies of the district occur in silicates, not in the ore suite. At Georgetown, where the Crinoidal is variable lithologically, Mn and Al gradients are not observed.

In the suite of specimens (Section C, Fig. 15) collected toward the magnetite-(garnet) mass, a moderate Al high in specimen 60 probably indicates a few small garnet centers just ahead of the main garnet front. Mn has been deposited over a somewhat greater area (Specimens 57–60), which is recognizable in the field by iron-oxide staining of the limestones.

2. The average of Zn values for each suite at Hanover (i.e., Sections D, E, F, G, H, and J, and specimens 111–117) is highest for specimens from the open pits (111–117). Section J, collected away from gossan under the Parting shale, gives the only regularly decreasing gradient of Zn concentration, and has the second highest average. Intrusion of a Tertiary post-ore dike into Section J did not disrupt the gradient. Sections D through H show erratic, greatly reduced Zn content and are not significant in indicating location of ore bodies.

Garnet specimens 66, 72, 78, and 97 form an excellent gradient of increasing Zn content toward the ore bodies, but this may be partly coincidental, for limestones immediately under the Parting shale (Specimens 67, 73, 79, 87, and 98) give a highly erratic succession of values.

Zn content of Georgetown specimens is low and rather uniform. There are three moderate Zn highs from Lone Mountain, two immediately under the Parting shale. This may indicate weak Zn mineralization like that throughout the rest of the Santa Rita district, in addition to the Ag mineralization in the Fusselman.

3. The two highest Ag values (29 and 40) are both from Lone Mountain. In general, Ag content is so low and uniform as to be valueless for trace element study.

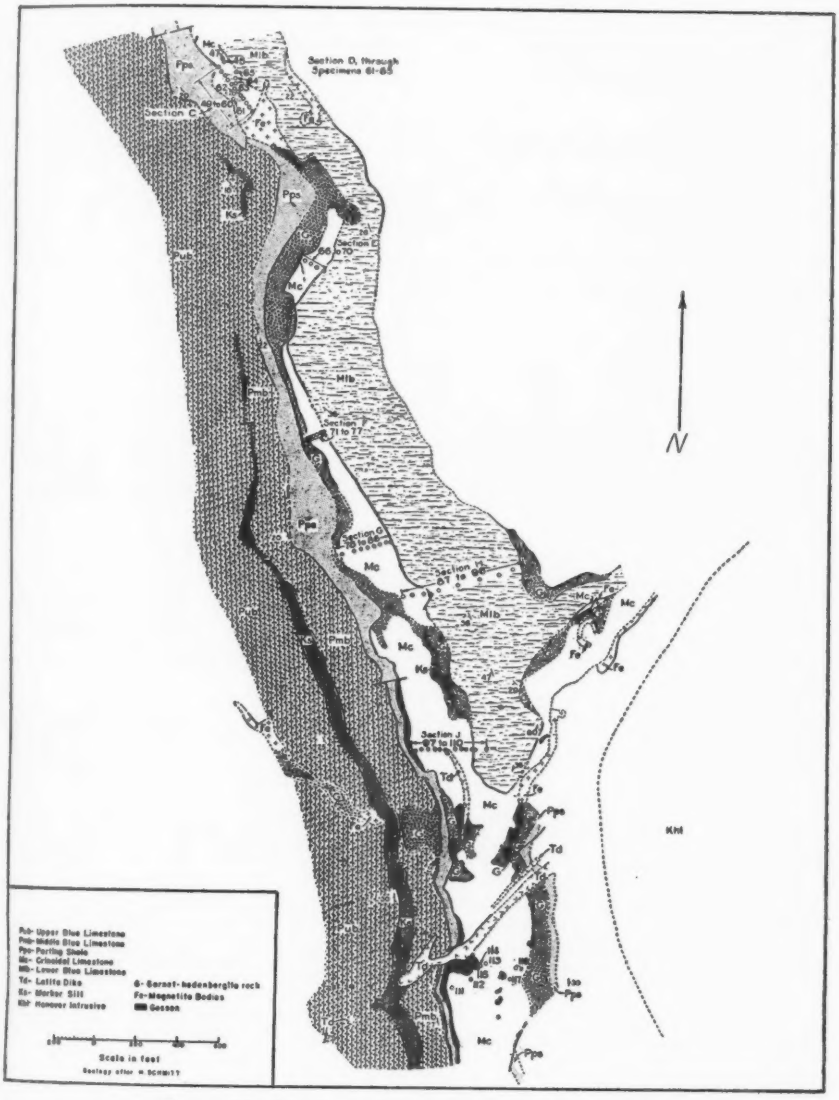

FIGURE 15.—LOCATION OF HANOVER SURFACE SPECIMENS

4. Average Cu values in Buckhorn Gulch are slightly greater in sections close to ore, but the difference is not sufficient be dependable.

5. Pb content of Hanover specimens in the open pits (Specimens 111–117) is about the same as that of other Buckhorn Gulch specimens. A simple spatial correlation between high Pb values and ore bodies does not exist; it is possible only to say that there are highs in Sections J, H, G, and F, while Sections D and E are uniformly low.

In individual specimen suites from Hanover, high Pb values occur some distance under the Crinoidal limestone-Parting shale contact

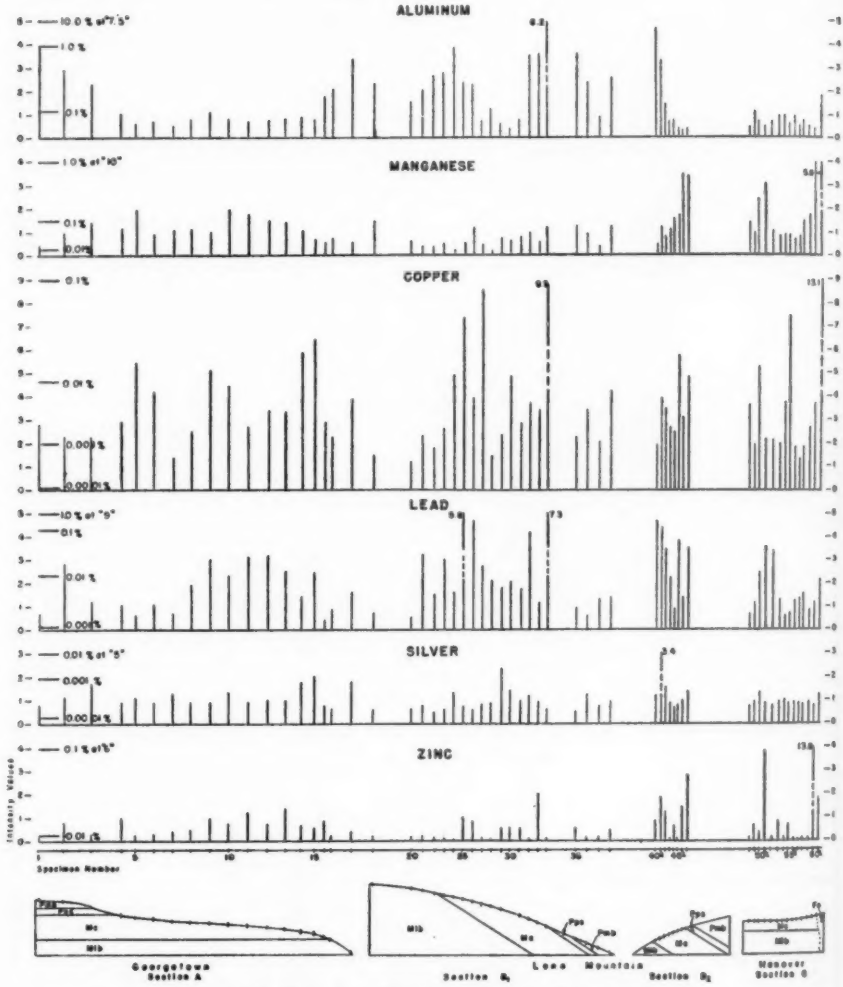

FIGURE 16.—DATA FOR SURFACE LIMESTONE SPECIMENS 1–60

Relative intensity values for different elements are not directly comparable.
Pmb—Middle Blue limestone
Pps—Parting shale
Mc—Crinoidal limestone
Mlb—Lower Blue limestone
Fe—Magnetite bodies

(Specimens 103 and 80, for example), apparently at an optimum balance between decreasing temperature and decreasing volume of solution flow. The Pb distribution is a small-scale repetition of the increase in Pb in ore bodies at increasing distances from Hanover.

These Pb highs are not sedimentary features, for at points distant from the open pit (Sections C, D, and E), the Crinoidal has a low, constant Pb content throughout.

Moderate Pb highs in the Georgetown specimens come from the middle Crinoidal lime-

stone, more fractured than the zone immediately under the Parting shale. Pb values from Lone Mountain are erratic in distribution, but their average is higher than that of even the most favorable suite of Hanover specimens.

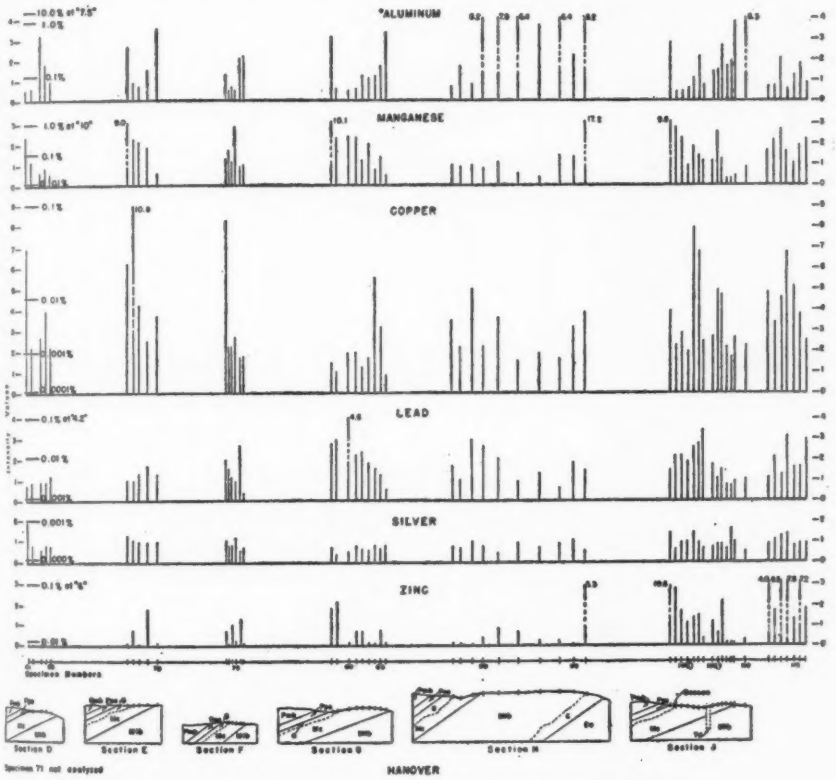

FIGURE 17.—DATA FOR SURFACE LIMESTONE SPECIMENS 61–117

Intensity values of different elements are not directly comparable.
Pmb—Middle Blue limestone
Pps—Parting shale
Mc—Crinoidal limestone
Mlb—Lower Blue limestone
Da—Augen member, Percha shale
Td—Tertiary quartz latite dike
G—Garnet-hedenbergite rock

Spectrographic Data, Underground Limestone Specimens

Underground specimens are from traverses at essentially constant stratigraphic horizons, proceeding away from dike-dault zones. Weathering and groundwater action is minimized in these specimens, obtained at depths from 100 to 1800 feet below the surface.

The separation of specimens on bar graphs (Figs. 19, 20) is not proportional to separation of points from which they were collected. Figure 18 is included to illustrate the typical geological situation and the exact location of specimens, but the conclusions which can be drawn from the data are not precise enough to justify including such maps for all specimens.

The occurrence of thin seams of pyrite-

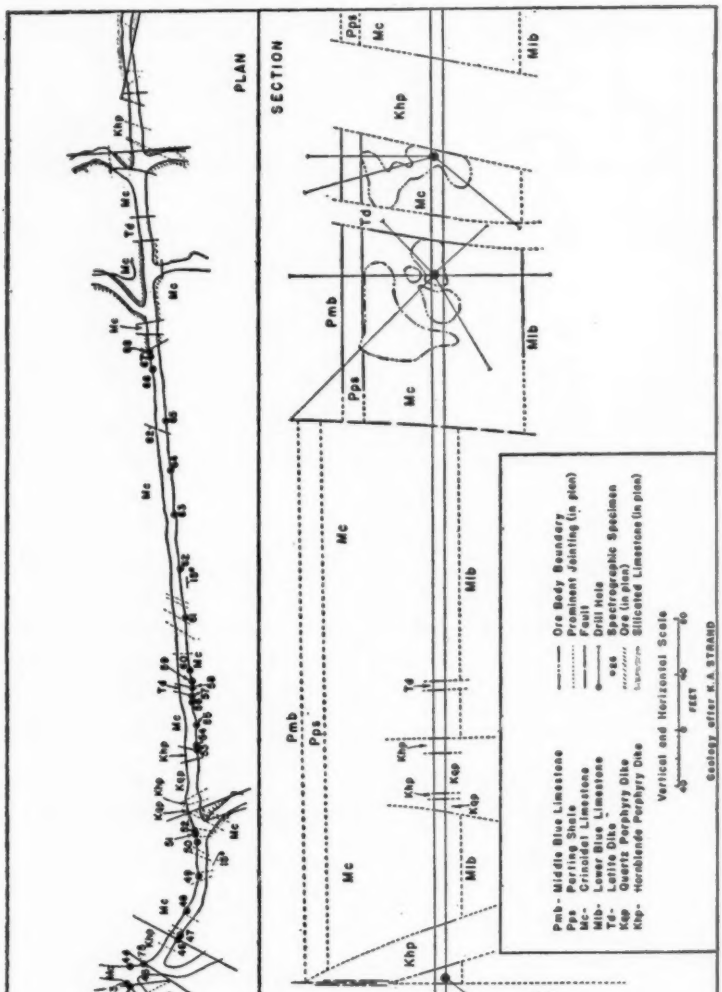

FIGURE 18.—SPECIMEN LOCATION MAP AND SECTION, OSWALDO MINE

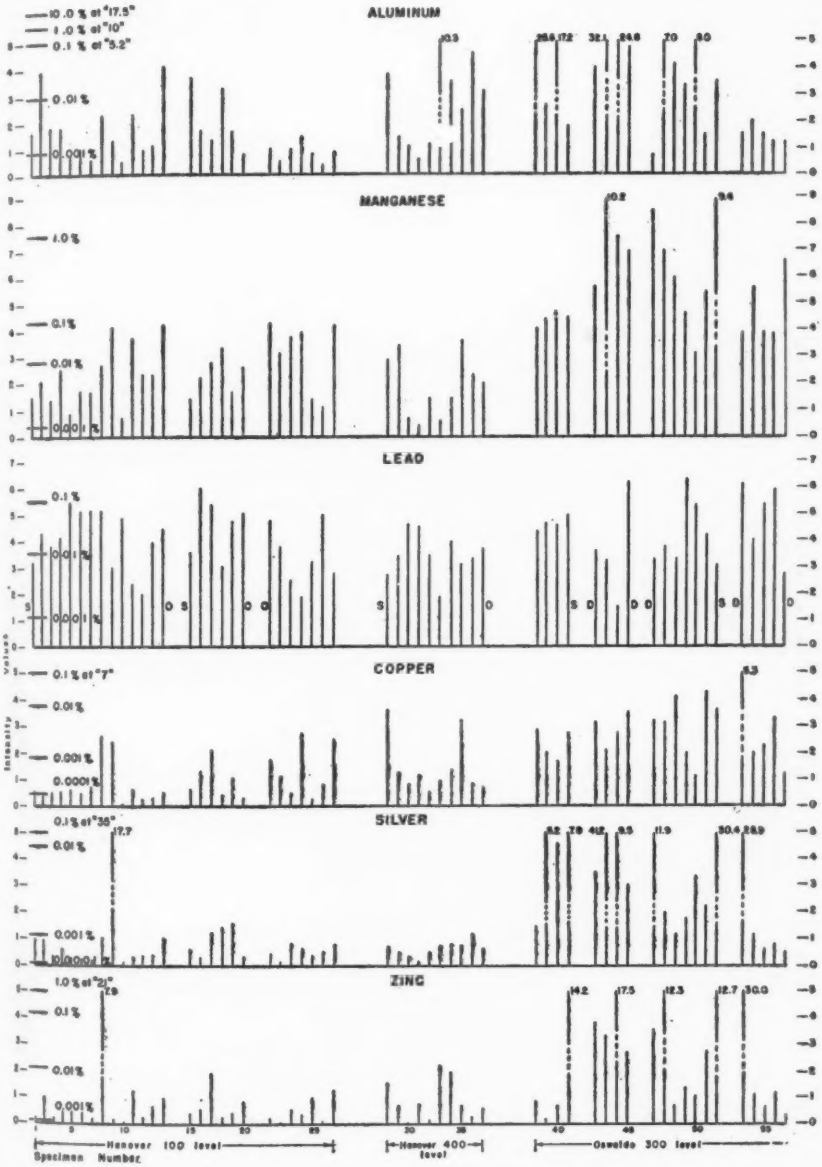

FIGURE 19.—DATA FOR UNDERGROUND LIMESTONE SPECIMENS 1–57

Relative intensity values for different elements are not directly comparable.
The symbols "O", "S", and "D" indicate, respectively, ore, silicates, and dike rock in the traverse.

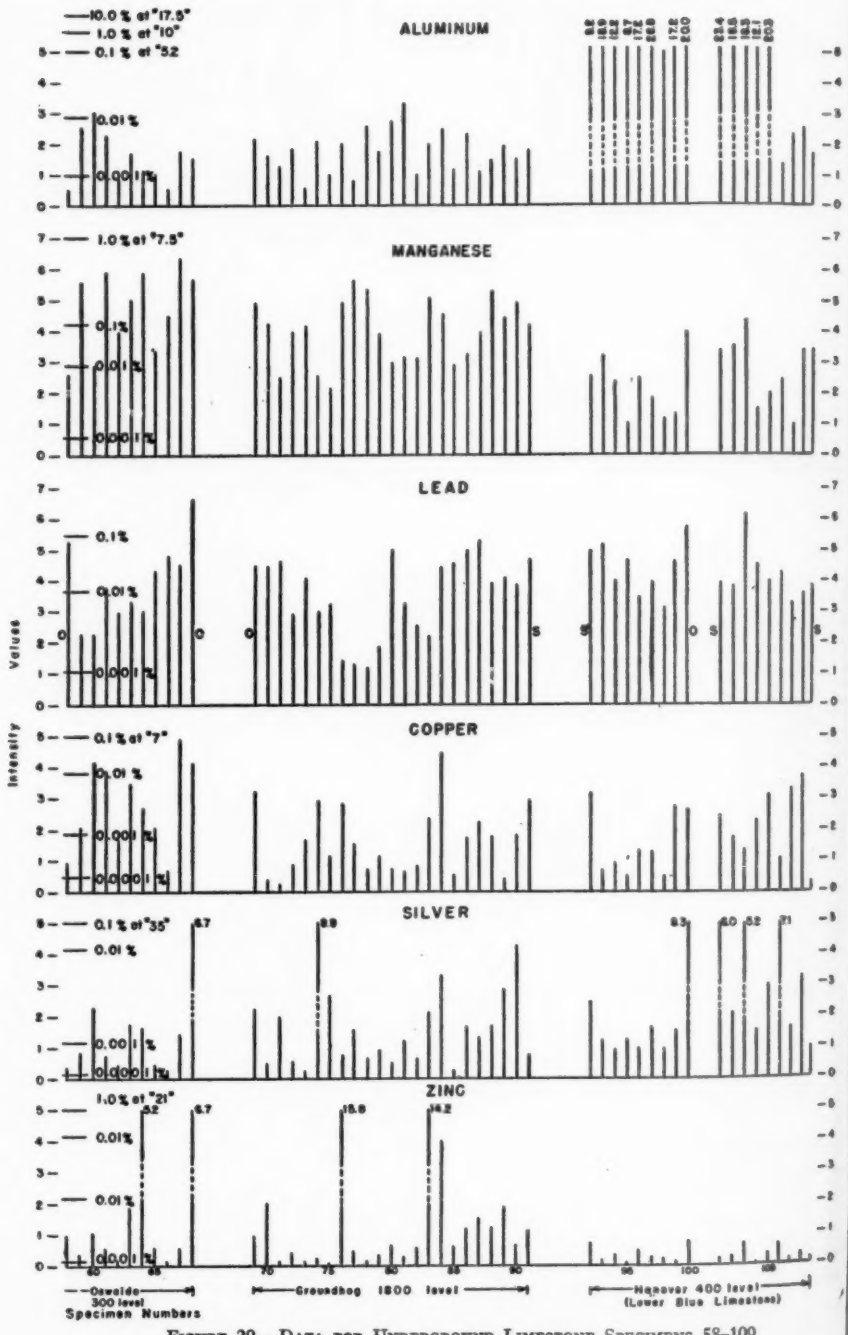

FIGURE 20.—DATA FOR UNDERGROUND LIMESTONE SPECIMENS 58–109

hedenbergite-calcite-sphalerite along fractures several score feet away from ore bodies, especially at the Groundhog mine, introduces a troublesome variable into interpretation of the outer portions of traverses.

These conclusions are based upon data in Figures 19 and 20:

1. Shaly Blue limestones give Al highs except for occasional calcitic layers. Erratic high Al values on the Hanover 400 level come from the lower Crinoidal, transitional into Lower Blue limestone. The only Al highs from silication lie within 2 feet of solid silicates (14, 15, 28, for example).

2. Cu highs lie both near ore bodies and dike margins, and farther away. They do not appear to be significant.

3. Ag is low at Hanover, but displays many highs from the badly broken and intruded Gamma dike-fault zone at the Oswaldo mine. Several rather strong Ag highs from Blue limestones at Hanover can hardly be sedimentational variations in such a limited area, in samples very similar lithologically. These highs, in an area where Crinoidal specimens contain very little Ag, suggest that the preference of Ag for impure limestones at Georgetown may apply also at Hanover.

4. Mn is highest near the Gamma dike-fault zone at the Oswaldo, but in general does not show a progressive change away from ore bodies and silicate zones.

5. Good Pb gradients do not extend throughout the whole distance between silicate zones, and it seems obvious that traverses have exceeded the limit of such gradients.

However, the Pb content in limestones next to "O" (Figs. 19, 20) generally decreases for 20 to 30 feet away from the ore body, as though increased near the channel by ore deposition. Limestone next to "S" is generally low in Pb at the channel, increasing to a peak within 20 or 30 feet. Pb introduced during silication, or removed from wall-rock at the channel, may have been redeposited at points of lower temperature. If this hypothesis is applied to surface limestone samples from Buckhorn Gulch, the conclusion is that the Pb distribution in individual traverses there is mainly controlled by silication rather than by ore desposition.

Conclusion

Pb is an ideal tracer element in all areas. In some places Zn has been leached, and in others it has been redistributed to give a magnified but useful gradient. Elements like V and Cd, whose minerals may stand out in oxidized outcrops, are not sufficiently abundant in the primary ore suite to be detectable in country rock above the background content. Ag, for the same reason, is significant only in Tertiary clays. Mn and Al were not deposited contemporaneously with ore at Hanover, and Mn is too widespread in the Tertiary to be useful.

Information obtained in the Tertiary area proves to be a contribution to exploration. The information obtainable at Hanover from conventional methods is unusually informative and the structure is quite complicated, hence improvements in sampling technique will be required before trace analysis can serve to add much to information now available as an adjunct to drilling. More detailed sampling will be needed at Georgetown and Lone Mountain, but present indications of the utility of the method at these localities are promising.

Trace gradients extend only a few feet from fractures, and larger-scale metal zoning results from superimposed gradients along many such fractures. Channel sampling, rather than collecting hand specimens, seems an effective way to overcome the localized influence of fracturing in trace element work. An effort would seem worthwhile to develop portable spectrographic units for use in local areas. The development of a spectrographic method that effectively represses selective volatilization would also seem desirable.

Trace element studies appear to represent a logical and more precise extension of present techniques for inspecting fault gouges, and searching for mineralization along faults and joints. The present study is only an introduction to what could very well be an integrated program, involving different observers at various localities differing in country rock, ore minerals, and physicochemical conditions of ore deposition.

These studies appear to offer a better chance of success if the following factors are applicable:

1. Extensive outcrops are available, giving a choice of sampling patterns.

2. The climate is arid or semi-arid, limiting the severity of leaching, and cover by vegetation.

3. The area is large enough and the objectives broad enough to permit many specimens to be examined. Errors in individual determinations because of sampling difficulties or spectrographic variables necessitate numerous confirmatory duplicate traverses. The effort required to develop spectrographic techniques is considerable, but once established, samples can be analyzed quickly and inexpensively.

4. The technique is made applicable to a wide variety of metals, and even non-metallics. The inclusiveness of spectrographic analysis, and the intensive prospecting required, make it desirable that maximum benefit be derived from a single study of each area.

5. The metals do not give weathered outcrops that are readily recognizable, or are commercially valuable at low metal content.

6. The geology is as little complicated as possible by repeated faulting, intrusion, and minor periods of mineralization.

7. A certain minimum of information is available concerning structural controls of ore deposition and geochemical suites of metals encountered in the broad area being studied. This should imply that chemical, spectrographic, or mineralogic examination of ore suites has been made at the nearest operating mines. In general, examination of material from within fractured zones may yield useful information even though the country rock is virtually impenetrable to ore-bearing solutions. Likewise, an altered rock may more often yield data of interest than a fresh one.

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Illinois Geological Survey, Urbana, Illinois; Columbia University, New York, N. Y.

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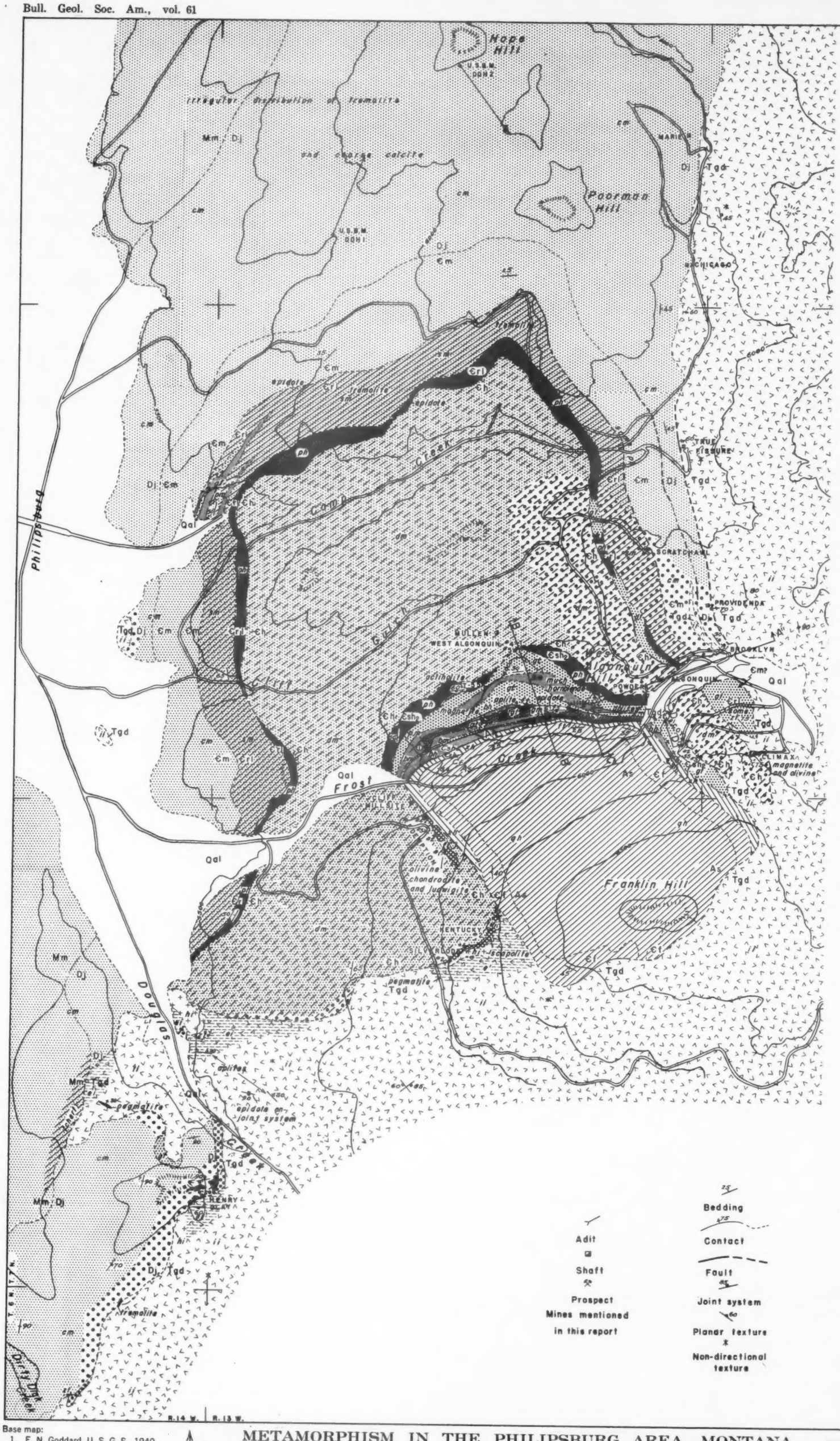

EXPLANATION

STRATIGRAPHY

Tert. Quat.	Qal	Alluvium
	Tgd	Philipsburg granodiorite
Miss.	Mm	Madison limestone
Dev.	Dj	Jefferson limestone
Upper Cambrian	€m	Maywood formation
	€rl	Red Lion formation
	€h	Hasmark dolomite
Middle Cambrian	€sh2	Silver Hill formation: Upper part
	€sh1	
	€f	Lower part
Belt ian	As	Flathead quartzite
		Spokane formation

LITHOLOGY

Mostly bleached and slightly recrystallized

- €m — Calcic limestone
- dm — Dolomitic limestone
- Sm — Siliceous banded calcic limestone
- Coarsely crystalline marble (>1mm)
- bm — Biotite-calcite banded marble
- pm — Diopside-calcite banded marble
- Dml — Diopside-grossularite-calcite banded marble
- gt — Diopside-grandite banded tactite

some hornblende

- Zoned garnet-epidote tactite
- ml — Magnetite-hornblende banded tactite (some garnet and pyroxene)
- Magnetite
- bh — Biotite-orthoclase-quartz banded hornfels (some pyroxene)
- oh — Undifferentiated siliceous hornfels
- Cordierite-biotite spotted hornfels
- qh — Quartzite with interbedded hornfels
- Poikilitic calcite-actinolite rock (some scapolite)
- Normal granodiorite
- Epidotized granodiorite
- NL — Amphibolitic intrusive border

Adit

Shaft

Prospect

Mines mentioned in this report

Bedding

Contact

Fault

Joint system

Planar texture

Non-directional texture

METAMORPHISM IN THE PHILIPSBURG AREA, MONTANA

Base map:
1. E. N. Goddard, U. S. G. S., 1940
2. W. T. Holser, assisted by J. W. Allingham, 1947

0 500 1000 2000 FEET

CONTOUR INTERVAL 250 FEET

Stratigraphic geology:
1. E. N. Goddard, U. S. G. S., 1940
2. W. T. Holser, 1947

Metamorphic geology:
W. T. Holser, 1947-1948.

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METAMORPHISM AND ASSOCIATED MINERALIZATION IN THE PHILIPSBURG REGION, MONTANA

By William T. Holser

CONTENTS

	Page
Abstract	1054
Introduction	1054
General geology	1056
Introduction	1056
Paleozoic sediments	1056
Granodiorites	1056
Field relations in the metamorphic areas	1057
General statement	1057
Philipsburg area	1059
General statement	1059
Northern part	1059
West Algonquin Hill	1060
Southern part	1061
Other areas bordering the Philipsburg batholith	1062
Southern Cross-Cable area	1063
General statement	1063
Cable mine	1065
Daly Gulch area	1065
Olson Gulch area	1067
Garnet area	1067
Isochemical metamorphism	1067
Macrocrystallization and chemical composition	1067
Marmorization in relation to deformation	1068
Fracture control of marmorization	1069
Bleaching of the Madison limestone	1069
Calcium-deficient rocks	1069
Impure limestones: isochemical reaction marbles	1071
Allochemical silicate metamorphism	1072
Grossularite marbles	1072
Silicate tactites	1073
Magnetite mineralization	1074
Magnetite tactite	1074
Hornblende in allochemical metamorphic rocks	1074
Magnetite-carbonate marble and related rocks	1075
Control of magnetite deposition, and relation of magnetite tactite to magnetite-carbonate marble	1076
Chemistry of magnetite deposition	1077
Late magnetite veins	1080
Aplite sills	1081
Occurrence	1081
Composition	1081
Endomorphism of the intrusives	1082
Zoned-garnet tactite	1082
General features	1082
Mineralogy	1082
Scapolitization	1083
Areal distribution	1083
Structural control	1083
Paragenesis	1084
Origin	1086
Joint-plane mineralization	1086
General features	1086
Mineralogy	1086
Chemical origin	1086
Structural origin	1087
Related occurrences	1088
Sulfide and related mineralization	1088
References cited	1089

ILLUSTRATIONS

Figure	Page
1. Index map of Cretaceous-Tertiary intrusives in parts of Montana and Idaho	1055
2. Metamorphism in the Philipsburg region, Montana	1057
3. Geology of tunnel no. 4, Silver Lode mine, Philipsburg area	1062
4. Geology of the Pomeroy mine, Southern Cross-Cable area	1064
5. Details of metamorphism at contact of Cable stock, Daly Gulch area	1066
6. Details of metamorphism at West Algonquin Hill, Philipsburg area: Silver Hill formation in vicinity of northeasterly fracture: Geologic relations	1070
7. Details of metamorphism at West Algonquin Hill, Philipsburg area: Silver Hill formation in vicinity of northeasterly fracture: Chemical relations	1070

Plate	Facing page
1. Metamorphism in the Philipsburg area	1053
2. Details of metamorphism at West Algonquin Hill, Philipsburg area: Sections perpendicular to axis of anticline	1060
3. Details of metamorphism at West Algonquin Hill, Philipsburg area: Sections parallel to axis of anticline	1062
4. Geology of the Cable mine, Southern Cross-Cable area	1066
5. Details of metamorphism at West Algonquin Hill, Philipsburg area: Silver Hill formation in vicinity of northeasterly fracture: Mineralogic relations	1070
	Following page
6. Areas of intense metamorphism 7. Garnetization of impure limestones 8. Mineralization in the Cable mine 9. Magnetite mineralization	1078

Following page

10. Intrusive rocks
11. Scapolitization
12. Zoned-garnet tactite
13. Late mineralization 1084

Facing page

14. Metamorphism in the Southern Cross-Cable area 1090

TABLES

Table	Page
1. Character of unmetamorphosed Paleozoic sediments	1058
2. Partial chemical analyses of carbonate rocks, Philipsburg region	1068
3. Chemical analyses of metamorphosed impure limestones from the Silver Hill formation, West Algonquin Hill, Philipsburg area	1071
4. Chemical analysis of amphibole from Philipsburg, Montana	1075
5. Chemical analysis of magnetite-hedenbergite tactite, Philipsburg area	1078
6. Chemical analysis of artificial hornblende	1078
7. Relative abundance of minerals filling joints in various types of rocks	1087

Abstract

Intrusive contacts of several granodiorite bodies, with calcitic, dolomitic and shaly limestones are described in detail. Isochemical metamorphism included recrystallization and formation of minerals of the pyroxene hornfels facies; it was succeeded by several phases of alteration with addition of material. In the shaly and silicic limestones, addition of iron and smaller quantities of silica and magnesia altered biotite-calcite marble, through several phases, to a banded andradite-diopside tactite. Fluo-hydrothermal introduction of much iron subsequently altered this tactite to one composed predominantly of magnetite and pargasite; simultaneously magnetite and forsterite replaced the pure carbonate formations. Coarse, concentrically zoned garnet was deposited at some intrusive contacts. In the succeeding phase, soda-rich hydrothermal solutions locally replaced nearly all types of rock with scapolite or calcite; the iron was reconcentrated as magnetite at some places in the scapolite. In a final phase, epidote, andradite, and other minerals, derived from the surrounding rock, filled the joints. Pyrite, then other sulfides and gold, also locally followed the scapolitization.

Fractures controlled all phases of the metamorphism, except probably the early reconstitution, although metamorphism was generally more pronounced near intrusive contacts. Most of the controlling structures are older than the intrusives. In the shaly limestone at Philipsburg, thin aplites were emplaced along the bedding by a combination of replacement and stoping.

Introduction

This study is part of a continuing investigation of allochemical metamorphism and its relations to high-temperature hydrothermal activity. A detailed field study seemed desirable as a guide to later experimental and theoretical studies. The greatest need was for mapping on a scale large enough to detect local structural control and mineralogical variation. Many areas were considered, in the United States and Mexico, which showed both silicate metamorphism and magnetite or similar mineralization. A region near Philipsburg, western Montana (Fig. 1), was chosen for study, because of the variety of metamorphism and mineralization, the good exposures, known regional geology, and accessibility. Mr. E. N. Goddard suggested the region.

Five months of 1947–1948 were devoted to detailed geologic mapping of favorable areas; 200 thin sections and 25 polished sections were studied. A copy of the complete report is available in the libraries of Columbia University and Cornell University; specimens and sections are deposited in the petrographic collection of Cornell University.

The field work in both years was generously supported by the Special Research Fund in Economic Geology, and in 1948 also by the Kemp Memorial Fund, both of Columbia University. The writer was assisted during two months in 1947 by Mr. J. W. Allingham, Jr., of the California Institute of Technology, who did much of the mapping around the Cable stock. Thanks are due a large number of people in the Philipsburg region who helped with suggestions, interest, and general hospitality. In particular, Mr. L. V. Manning of the Trout Mining Division, American Machine and Metals Co., Mr. A. V. Taylor, Jr., of the Taylor-Knapp Mining Co., and Mr. R. R. Wallace of the Canyon Lode Mining Co., con-

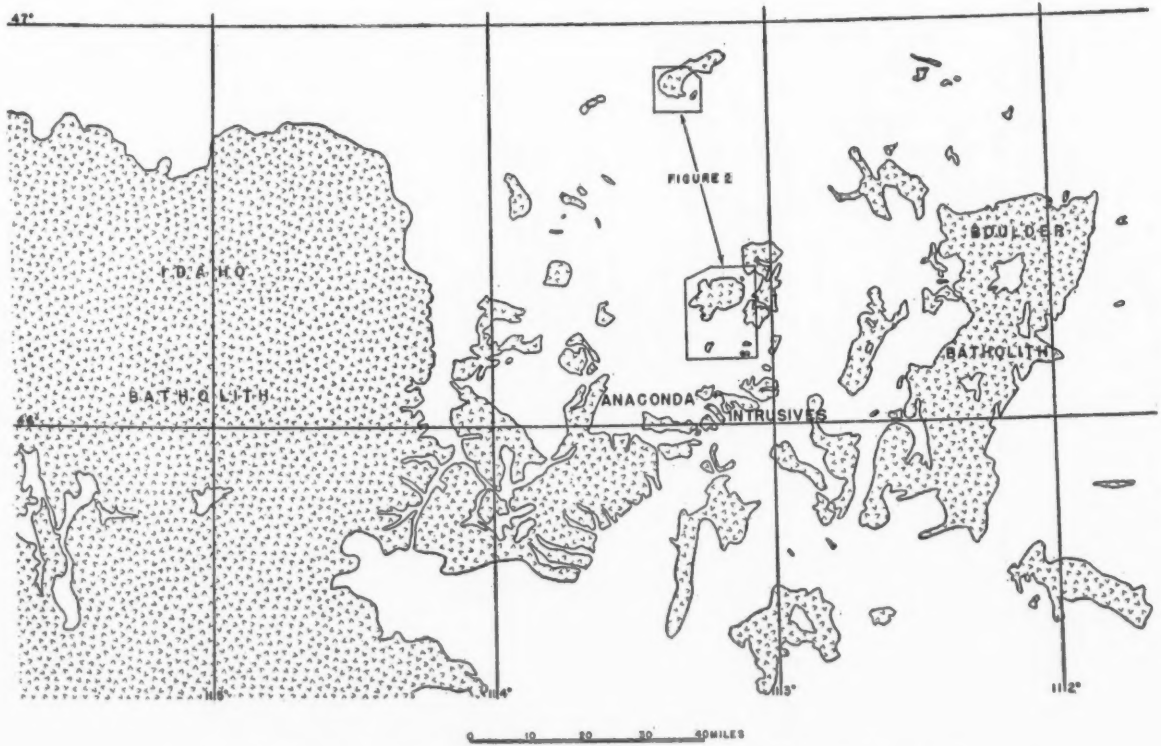

FIGURE 1.—INDEX MAP of CRETACEOUS-TERTIARY INTRUSIVES IN PARTS OF MONTANA AND IDAHO

tributed generously of their time and facilities and made available a number of private maps and reports.

The problem was undertaken at the suggestion of Dr. C. H. Behre, Jr., and his interest has been enthusiastic and inspiring. I appreciate the courtesy of Dr. S. J. Shand, Dr. F. J. Turner, and Dr. R. J. Holmes, who read the manuscript and contributed pertinent suggestions. My indebtedness to the many workers in this field is not reflected in the brief text references.

General Geology

Introduction

The geology of the Philipsburg quadrangle was well described by Emmons and Calkins (1913); and the Garnet Range was mapped by Pardee (1917). General features of ore genesis were discussed by Billingsley and Grimes (1917). In general a series of conformable Paleozoic calcareous sediments unconformably underlies Jurassic and Cretaceous calcareous sediments and Tertiary gravels and overlies, with regional unconformity, a thick quartzitic section of the Belt series (Fig. 2). While the most important faulting is thrusting from the west, a complex system of block faults appears locally. The sediments are folded rather steeply, parallel to the thrust strike, and intruded by many Tertiary stocks.

In order to limit the problem and control the conditions, the investigation was confined to the effects of one type of intrusive rock on a limited section of sediments: granodiorite, on the Cambrian formations and (to a lesser degree) on the Devonian and Mississippian limestones.

Paleozoic Sediments

Table 1 is a composite of measurements in various parts of the region by Calkins (Emmons and Calkins, 1913, p. 49–74), Pardee (1917, p. 167–169), Christina Lochman (personal communication, 1948), and the author. Petrography of unmetamorphosed Montana sediments is described in detail by Tweto (1937).

The common procedure of exact correlation of metamorphosed and unmetamorphosed sections could not be followed in most parts of this region. Some formations are exposed only for a limited distance from the intrusives, and formations such as the Silver Hill cannot be found completely unmetamorphosed except several miles from intrusions. The field work was primarily concerned with local variations in composition of thin, uniform beds.

Granodiorites

Figure 1 indicates the central relation of the post-Laramide intrusions of this region to the Idaho and Boulder batholiths, and those of the Anaconda Range. These intrusives vary locally around a mean composition of granodiorite; in the Philipsburg region the variation is from diorite to muscovite granite. The large intrusive bodies are all granitic; those of granodiorite are of interest in this initial study, including the Philipsburg and Garnet batholiths and smaller bodies at Cable, Olson Gulch, and Top O'Deep.

The selected group is uniform in general composition; all are medium-grained, biotite-hornblende granodiorites, with andesine and orthoclase. The plagioclase crystals are commonly zoned, with highly calcic cores and oligoclase rims. Chemical analyses are given by Calkins, with detailed petrographic descriptions (Emmons and Calkins, 1913, p. 94–104; *see also* Pardee, 1917, p. 169).

Each intrusive mass exhibits no more variation within itself than there is among the separate masses. The borders are richer in ferromagnesian minerals or show a concentration of lamprophyric inclusions, but only within a few feet of the contact may these minerals exhibit a recognizable planar structure, parallel to the usually steep contact. Contacts seen are very sharp, in some cases faulted, and locally discordant with the bedding. On a larger scale they are concordant with the bedding in some areas (Fig. 2, eastern half Philipsburg batholith); but in certain cases this concordance may have been induced by intrusive forces, as around the Cable stock (Fig. 2; Pl. 14; Pl. 6, fig. 2). At only one locality was any "granitized" border rock recognized. Satellitic intrusions are uncommon around the main intrusions. At a few places, as in the Philipsburg and Southern Cross areas, a dike of

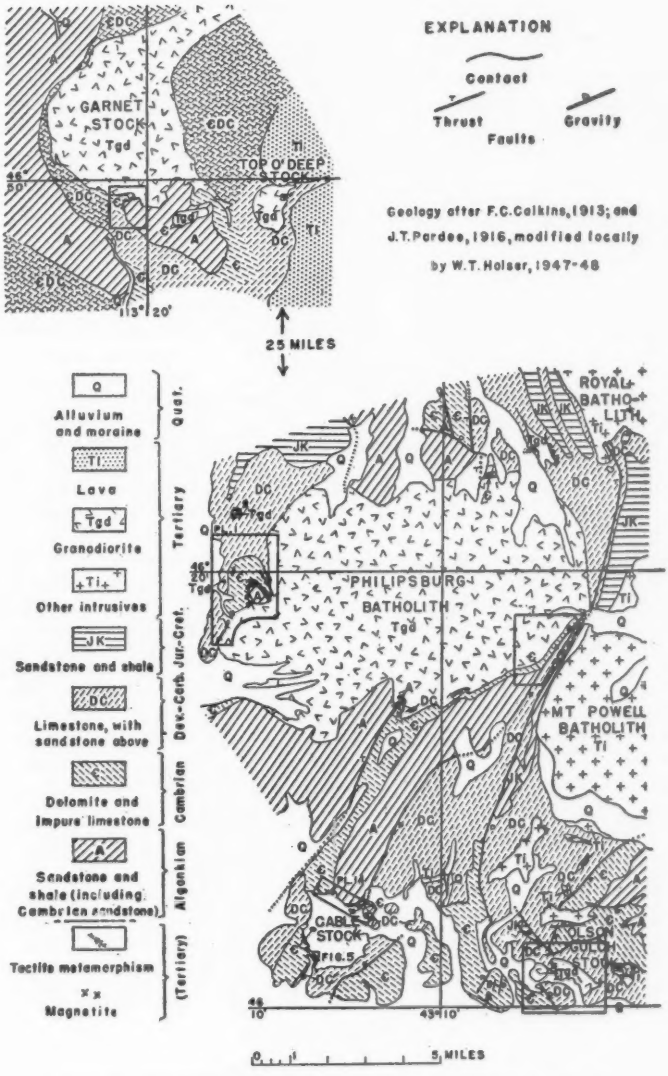

FIGURE 2.—METAMORPHISM IN THE PHILIPSBURG REGION, MONTANA

granodiorite projects from the main mass. Pegmatitic rocks are unusually rare, and aplite sills are quite localized in the Philipsburg area.

The total load at the time of intrusion was not more than 6000–8000 feet of sediments, or about 750 atmospheres.

FIELD RELATIONS IN THE METAMORPHIC AREAS

General Statement

In the Philipsburg region nearly all sedimentary rocks close to intrusive rocks show

TABLE 1.—CHARACTER OF UNMETAMORPHOSED PALEOZOIC SEDIMENTS

	Age*	Formation*	Thickness (Feet)	Character
Mississippian	Kinderhookian-Osagian	Madison limestone	1000–1500	Limestone, dark gray, aphanitic, thick-bedded to massive; calcitic, commonly with beds or large masses of chert, and petroliferous; lower 200 feet locally silty.
Devonian	Senecan	Jefferson limestone	1000–1300	Limestone, gray to brownish gray, aphanitic, massive; calcitic to slightly dolomitic, petroliferous.
Upper Cambrian	Trempaleauan	Maywood formation	200–450	Variable in composition: north of Philipsburg batholith buff to red fine-grained thin-bedded dolomitic limestone and siltstone, and gray-green to buff sandstone, in units 1–5 feet thick; south of the batholith buff fine-grained moderately thin-bedded slightly dolomitic limestone, with some sandy beds.
	Franconian	Red Lion formation	225–350	Limestone, light to dark blue gray, aphanitic to very fine-grained, calcitic; laminae of light-brown aphanitic siliceous shale or silica, which weathers dark red brown. The laminae and the limestone beds both vary in thickness from 0.1 to 1 inch; the proportion of siliceous material in the bulk rock ranges from 5 to 60 per cent, increasing downward; the laminae may be straight or wavy when thin, often broken; when in large proportion their anastomosis leaves the limestone as lenses (see Pl. 7, fig. 1). At base 0–20 feet of light-gray very fine-grained well-bedded calcic sandstone overlies 0–3 feet black shale.
	Dresbachian	Hasmark dolomite	Average about 900, but thinning to 350 near intrusives.	Limestone, light blue gray to white, weathering buff to white, aphanitic to fine grained, massive; dolomitic.
Middle Cambrian	(*cf.* Park shale-Meagher limestone†)	Upper part Silver Hill formation	150–250	Limestone with siliceous laminae, as in Red Lion formation, but laminae usually more shaly; bulk proportion of limestone decreasing upward to about 25 per cent, but varying widely both vertically and laterally.

TABLE 1.—*Continued*

	Age*	Formation*	Thickness (Feet)	Character
Middle Cambrian	(*cf.* Wolsey shale†)	Lower part Silver Hill formation	60–120	Shale, dark green to dark gray or violet, very thin-bedded; thin interbeds of white to maroon fine- to medium-grained sandstone, commonly glauconitic.
		Flathead quartzite	135–150	Quartzite, white to red, medium- to fine-grained, thick-bedded; partings and thin interbeds of green shale. Most exposures transitional downward into Beltian Spokane quartzites.

* Stratigraphy according to Calkins (Emmons and Calkins 1913) as modified by Lochman and Duncan (1944, p. 12–16).

† Upper part Silver Hill formation is the upper part and middle part of the formation as designated by Calkins (Emmons and Calkins 1913, p. 53–56), and by Lochman and Duncan (1944, p. 15); lower part Silver Hill formation is lower part of the formation as delimited by Calkins. All three parts mapped as one unit by Calkins, who states, "The highest and middle members are not distinct..."

some metamorphism, but in certain localities the metamorphism is intensely developed (Fig. 2). The best exposures of rocks showing the metamorphism were examined in detail. In addition to those illustrating this paper, large-scale maps were made in the Twin Peaks, Boulder Creek, Olson Gulch, Garnet, and other areas.

Philipsburg Area

General Statement.—A northwardly pitching Paleozoic anticline is covered on the west by alluvium and intruded in the east and southeast by the Philipsburg batholith (Pl. 1).[1] The limestones have been partially recrystallized, except for some in the northwest. In addition, certain areas show an extensive development of garnet, diopside, scapolite, magnetite, and other minerals. The rocks are described from northeast to southwest, along the intrusive contact.

Northern part.—Jefferson and Madison limestones in contact with the granodiorite are simply recrystallized. At well-exposed contacts, calcite grains scarcely exceed 0.05 mm. in diameter, with some original color retained. Tremolite, as well as epidote, phlogopite, and coarse calcite, are also found in certain beds, widely but very sparsely distributed. The distribution bears no relation to visible intrusive contact.

The Maywood formation is argillaceous and siliceous limestone. Tremolite and diopside occur irregularly near the crest of the anticline. The formation is poorly exposed at its contact with the granodiorite dike east of the Algonquin shaft, but no evidence of extensive silication was found there, indicating that something other than nearness to an intrusive contact is required for the development of hornfels, and especially of tactite, in impure limestones.

In the Red Lion formation, calcite is recrystallized to at least 0.05 mm. everywhere on the anticline. Thin sections of the siliceous or argillaceous laminae reveal small biotite or tremolite crystals. The lowermost 20 feet, corresponding to a bed of calcic sandstone in the unmetamorphosed section, is commonly metamorphosed to a hard light-greenish hornfels, but near the head of Cliff Gulch it is a coarser tactite. Ordinary tactite has alternate bands of dark-brownish-red garnet and dark-green diopside, parallel to original bedding. South of Frost Creek, within 250 feet horizontally of the intru-

[1] For further details regarding the structure and the later mineralization of this area *see* Goddard (1940), and Emmons and Calkins (1913). Plate 1 of this paper may be superimposed on Goddard's Plate 26.

sive contact, the entire formation is tactite. Black andradite is the commonest mineral, characteristically showing concentric color zoning; diopside is abundant. Epidote, andradite, and some magnetite crystals cover small vuglike openings and vertical joint planes in the more massive garnet.

The Hasmark dolomite is remarkably uniform in composition and metamorphism. Recrystallization is widespread, but extensive observation gave only a rough correlation between crystal size and distance from intrusive contact; the areas of coarse carbonate are shown approximately in Plate 1. On the anticline axis the average diameter is 0.1 mm., while in the east limb it averages 1.0 mm. However, several local areas near the axis show crystallization up to 0.3 mm.; in a few cases these are northward-trending zones of shearing or close jointing. The total effect of recrystallization is a pure-white sugary dolomitic limestone; it weathers light brown and finally forms a coarse carbonate sand. Near the Coyle workings the dolomite contains small flakes of phlogopite; but neither thin sections nor gravity concentration revealed other silicates, even this close to the intrusive contact. In one small area, next to the Climax shaft, it is altered to a hard green diopsidic hornfels, with some coarse garnet, epidote, and magnetite. Granodiorite is exposed in the workings below, but complex shearing and serpentinization obscure the relations.

West Algonquin Hill.—The lower Cambrian Silver Hill formation has been metamorphosed throughout the arc of its exposure. In the west limb of the anticline, north of Frost Creek, the major unit is 50 feet of biotite-calcite banded marble, similar to slightly metamorphosed Red Lion. In some marbles biotite laminae are superseded by laminae of diopside or pargasite (magnesian hornblende), and in others the diopside laminae are bordered by small euhedral light-brown grossularite crystals (Pl. 7, fig. 3). In a few localities, both types are confined to the immediate vicinity of vertical fractures trending N. 25° E. across the bedding, with a pure garnet rock next to the fracture (Fig. 6).

Near the axis the biotite-calcite banded marble becomes subordinate to a thick section of garnet-diopside banded tactite and grossularite-diopside-calcite marble (Pl. 6, fig. 1). The tactite garnet is darker than the nearly pure grossularite of the marble, and refractive index-specific gravity measurements indicate 30–50 per cent andradite. It is called grandite, to distinguish it from nearly pure grossularite or andradite found elsewhere. At some localities, garnet is concentrated in certain beds, around thin garnet-filled joints that cross the bedding. Joints filled with epidote and other minerals, similar to those in the Red Lion, are very abundant; they trend easterly and dip steeply south.

The west base of the upper Silver Hill is light-greenish to reddish banded quartz-orthoclase hornfels, containing some biotite and a little tourmaline. Scapolite is concentrated in one area, adjacent to vertical N. 50° E. joints.

Everywhere on the anticline, the lower Silver Hill is a series of hard, dark hornfelses and quartzites. Most common is a very dark yellowish-gray cordierite hornfels with knots of biotite up to half an inch in diameter. Some are thinly banded with biotite, causing phyllitic fracture. Hornfels is interbedded with quartz-rich rocks, the bulk composition ranging to pure quartzite.

In the eastern limb of the anticline the relations are complex, and as they are well exposed in the cliffs along the north side of Frost Creek, west of the Algonquin shaft (Pl. 6, fig. 1), these were mapped on a large scale.[2] Tactites and garnet marbles on the cliffs predominate in sections corresponding to biotite marbles in the west. Much of the lower part of the formation is a magnetite tactite, with variable amounts of pargasite, garnet, and diopside-hedenbergite. The orthoclase-quartz hornfels and the biotite-calcite marble of the west limb are not seen. The magnetite tactite is dark greenish black, fine- to coarse-grained; and banded in a manner suggestive of the marble banding (Pl. 9, fig. 1).

An unusual amount of scapolite rock is concentrated near the base of the upper Silver Hill. Of two principal types, one is a light-colored rock of coarsely prismatic scapolite and finer-

[2] The cliffs are depicted in Plate 2,B as an elevation, projected to a vertical N. 70°E. plane, perpendicular to the axial plane. Plate 2,A locates the elevation. To emphasize structural irregularities within each outcrop area (bounded by a dotted line), all irregularities due to the roughness of separate surfaces of the outcrop are averaged out in projection

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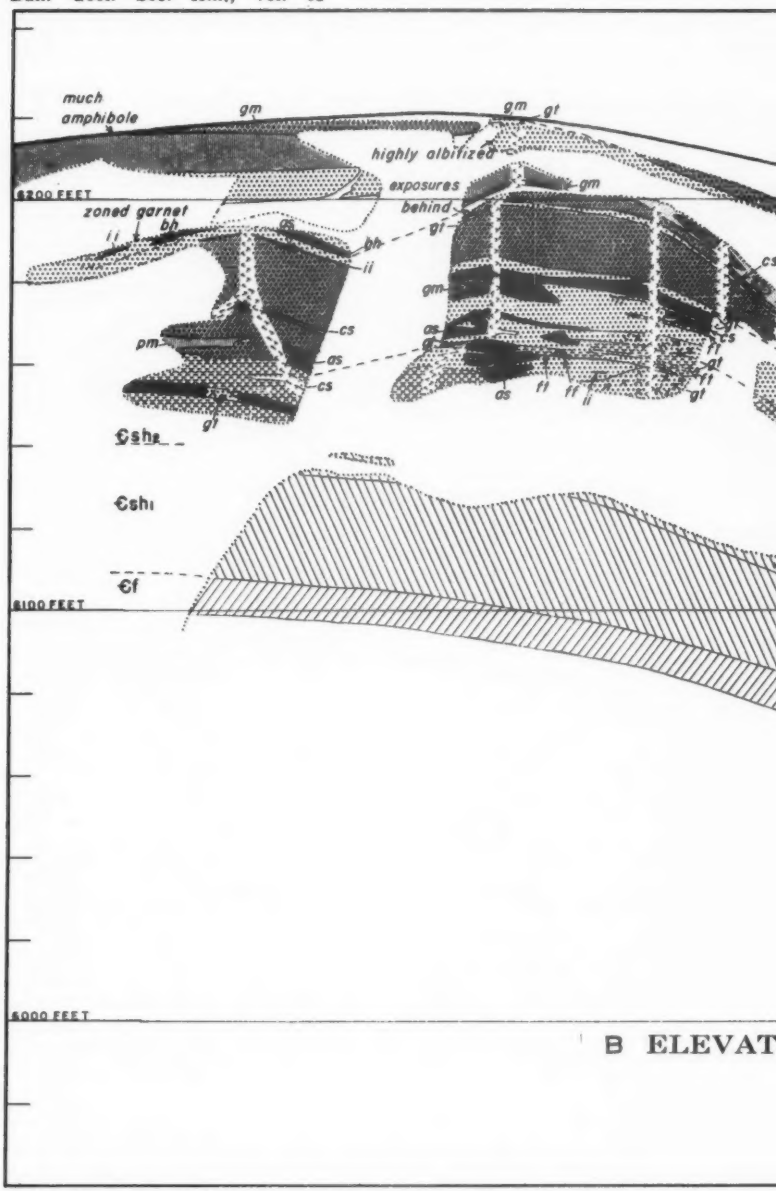
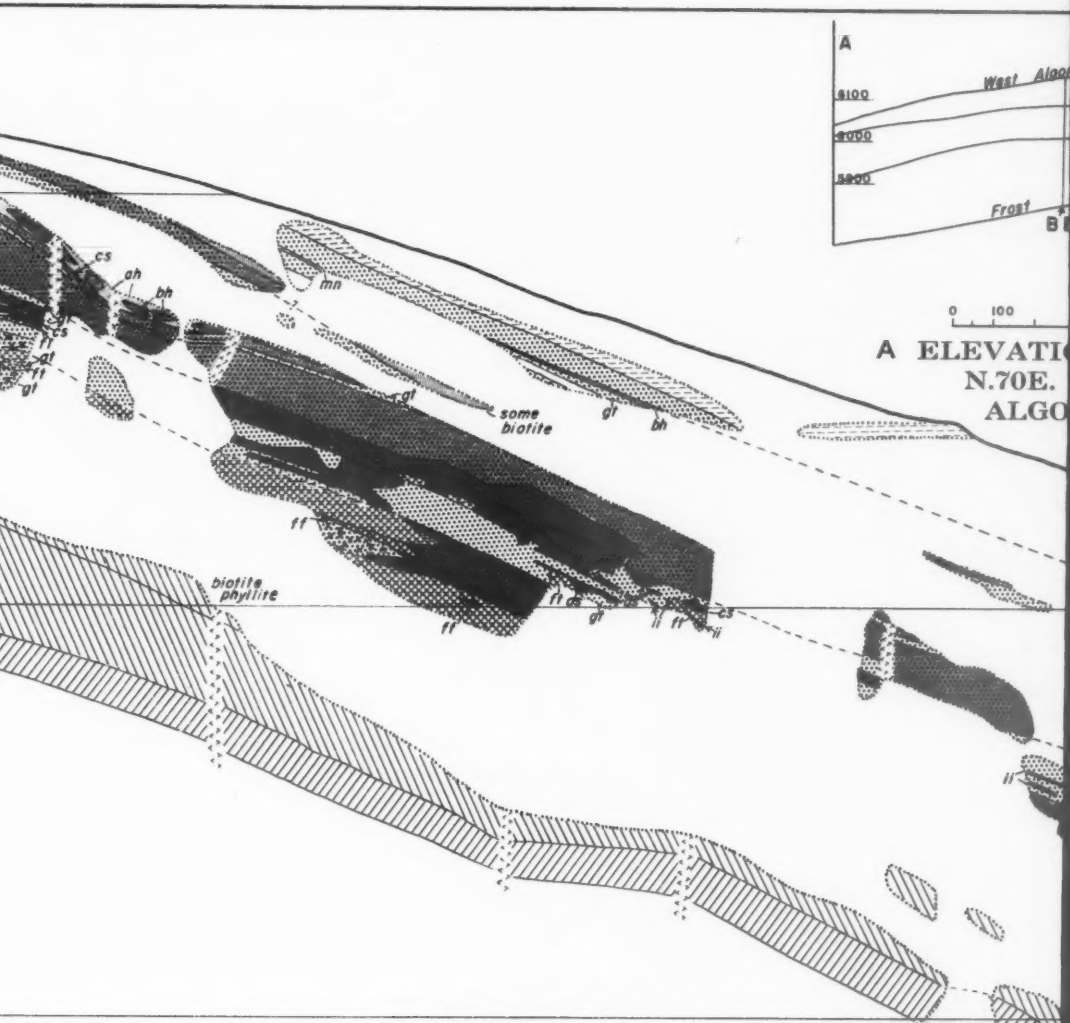


Projected to N.70E. Plane



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LEVATION OF PART OF WEST ALGONQUIN CLIFFS

Horizontal and vertical scale in plane of projection only

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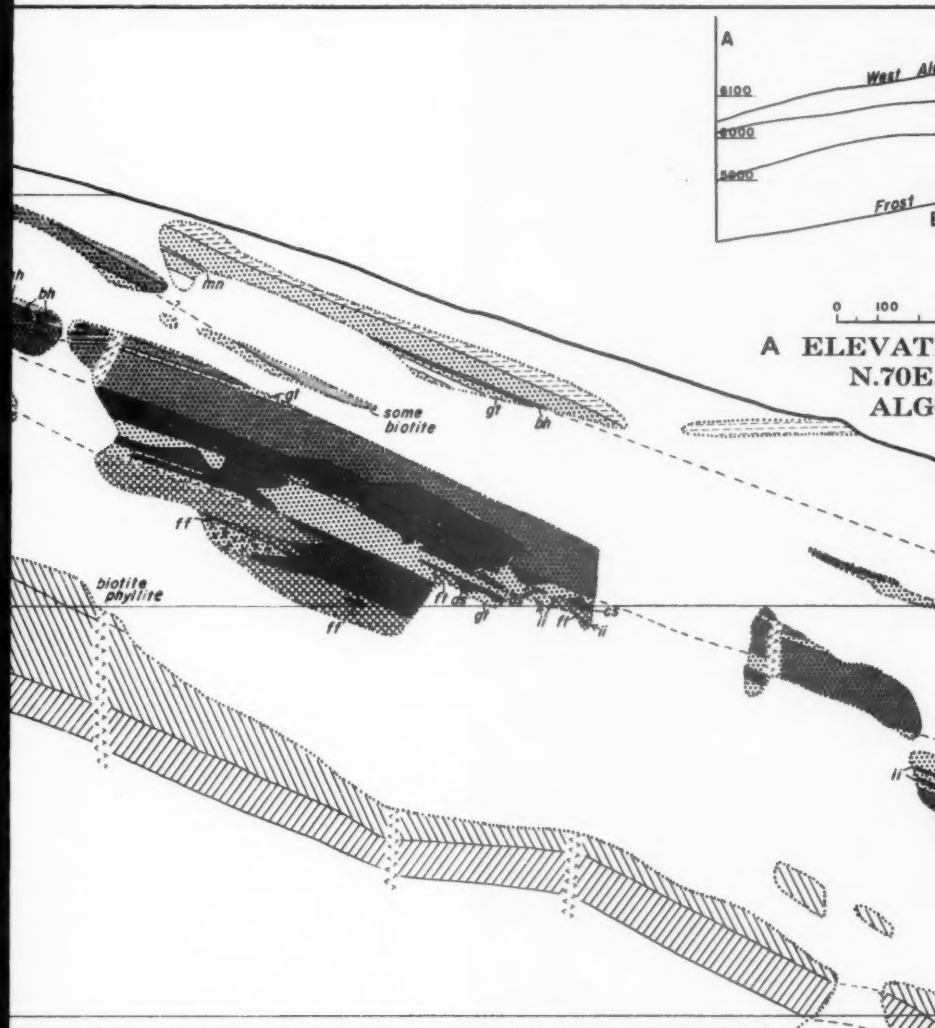
DETAILS OF METAMORPHISM AT WEST A

Sections Perpendicular



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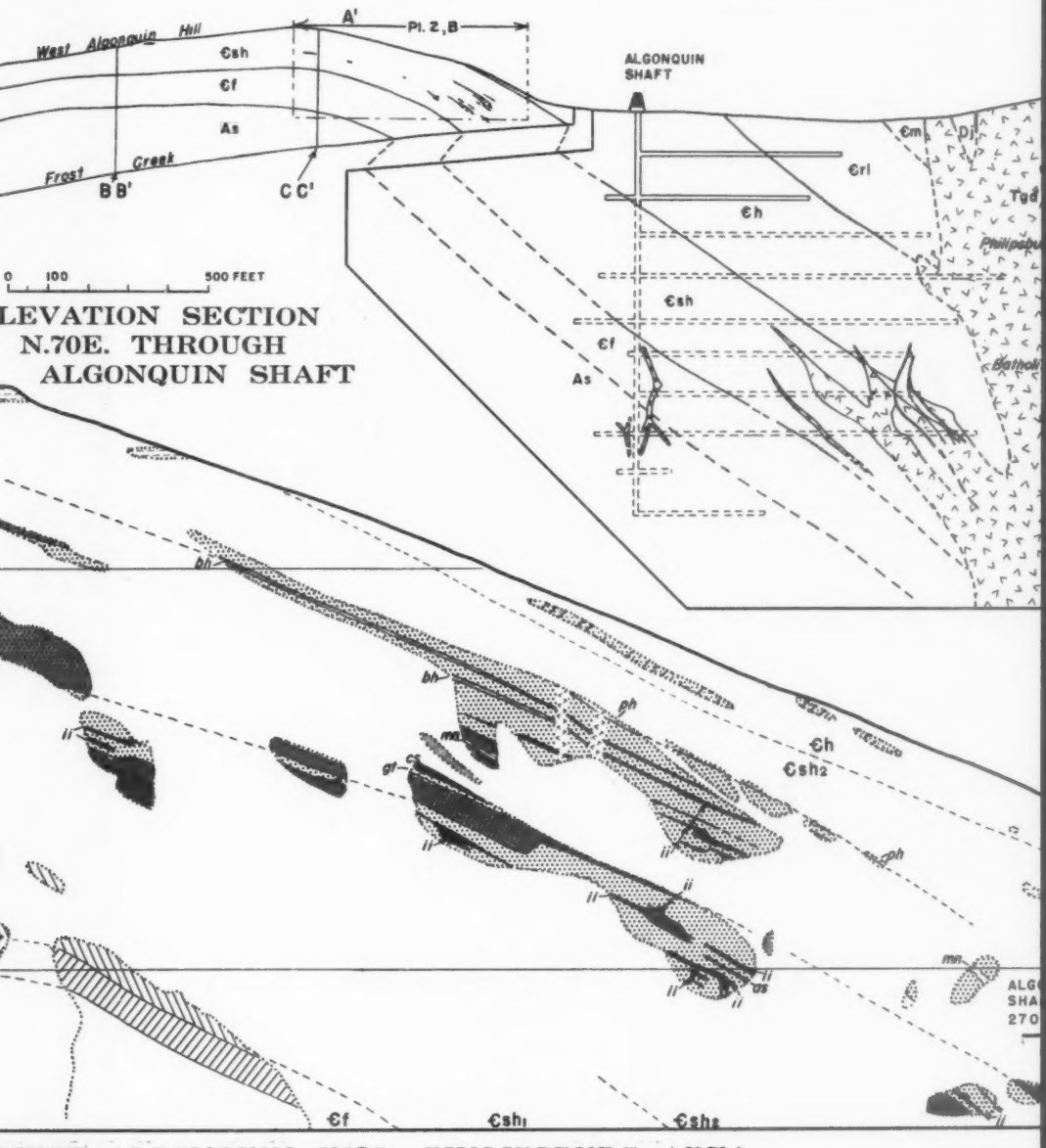

LEVATION SECTION
N.70E. THROUGH
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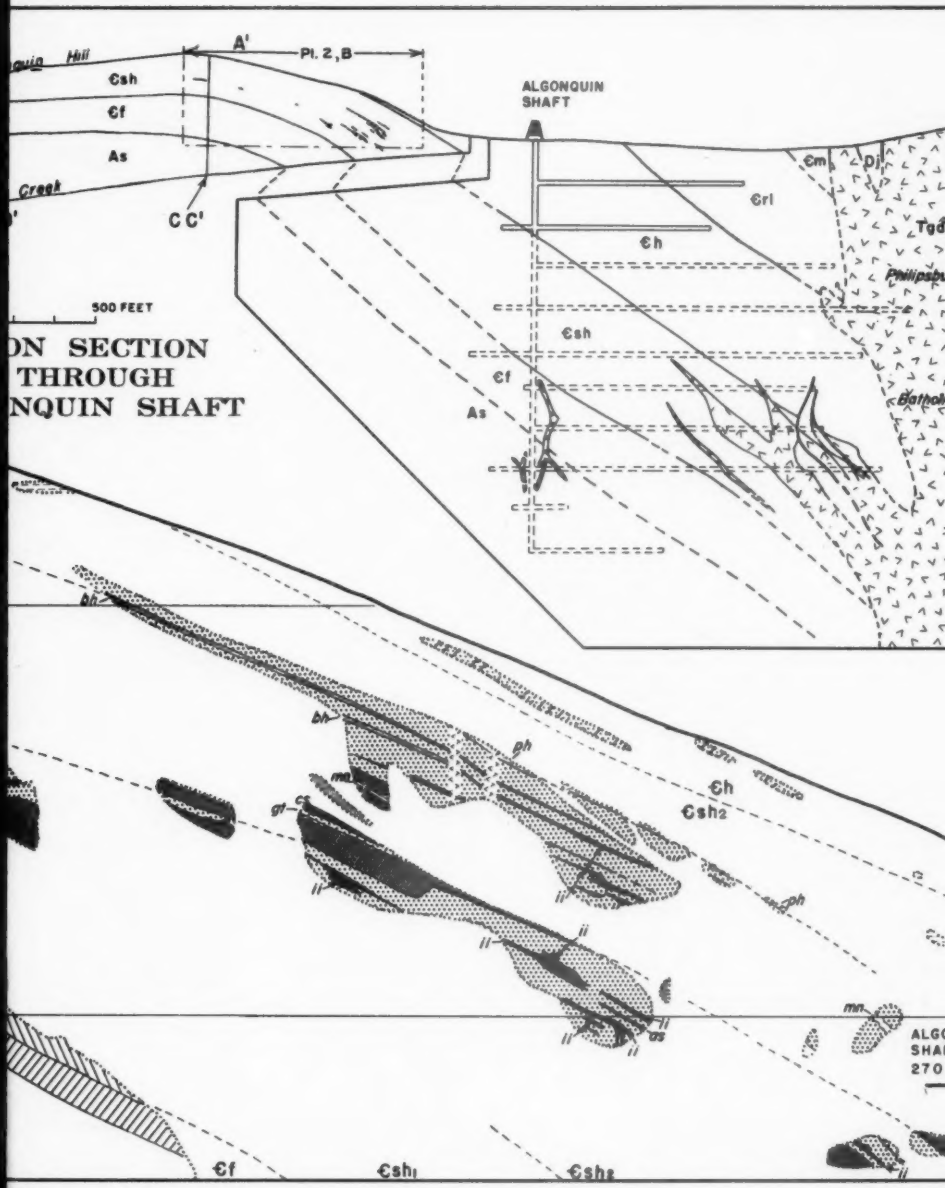
WEST ALGONQUIN HILL, PHILIPSBURG AREA

erpendicular to Axis of Anticline



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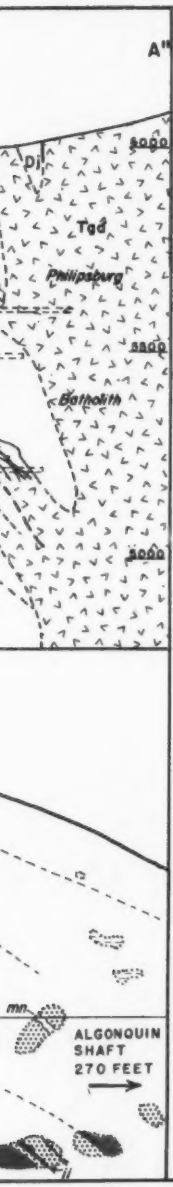

EXPLANATION

LITHOLOGY

- Dolomitic marble
- Biotite-calcite banded marble
- Diopside-calcite banded marble (some hornblende)
- Diopside-grossularite-calcite banded marble
- Diopside-grandite tactite
- Biotite-orthoclase–quartz hornfels (some pyroxene)
- Bleached hornfels
- Pyroxene hornfels
- Magnetite-hornblende banded tactite
- Magnetite
- Poikilitic calcite-actinolite rock (some scapolite)
- Scapolite-hornblende rock
- Pyrolusite-clay alteration
- Cordierite-biotite spotted hornfels
- Aplite
- Quartzite
- Outcrop
- Change in attitude of outcrop plane

STRATIGRAPHY

- Middle Cambrian
  - €h — Hasmark dolomite
  - Silver Hill formation
    - €shz — Upper part
    - €shl — Lower part
  - €f — Flathead quartzite
- Belt ian
  - As — Spokane formation
- Tert.
  - Philipsburg granodiorite

W. T. Holser, 1947
(Algonquin Mine below 200 level from maps of American Machine and Metals Co.)

grained very dark-green hornblende, with a variable amount of magnetite. The other is a dark rock of calcite in poikiloblastic crystals up to 2 inches in diameter, with numerous penetrating needles of light-green actinolite, some diopside, and little scapolite. These rocks irregularly crosscut the bedding (Pl. 2B; Pl. 11, fig. 1). Ovoid pods of pure scapolite, in prisms up to 6 inches long, occur in the scapolite rock; and similar areas in the poikiloblastic calcite rock are pure, coarsely crystalline calcite. Pods of pure magnetite occur locally in scapolite rock, and much of the hornblende-magnetite banded tactite contains some scapolite.

Aplite sills are common in the east limb of the Silver Hill. Although most are sill-like, some branch into dikes; the borders are highly irregular. They contain rotated inclusions of the wall rock (Pl. 10, fig. 1) which is usually grandite-diopside tactite. In some places the aplite changes gradually along the outcrop to a coarsely crystalline scapolite rock, which irregularly penetrates the walls.

Section BB′ of Plate 3[3] indicates that the various metamorphic types are not very continuous down dip; the thick section of garnet, magnetite, and scapolite rocks of the cliff face is still seen in the highest tunnel but not in the lowest. Apparently the changes down the anticlinal axis are similar to the changes east to west around the outcrop.

Plate 2 A[4] indicates the proximity of the Algonquin Hill outcrops to the batholith; it further suggests a high susceptibility of the Silver Hill to intrusive action, much of it with faulted contacts. South of Frost Creek, several large dikes contact folded and brecciated grandite-diopside tactite of the Silver Hill (Pl. 3, sec. DD′). The intrusive rock is coarse-grained, dark-gray granodiorite with faint planar structure of the ferromagnesian minerals and many inclusions. The southern contact with the batholith is obscured by quartzite slide rock.

The Flathead and Spokane formations show only recrystallization of quartz and presence of biotite. They have no tactite, magnetite, or scapolite mineralization and were not investigated in detail. Along the western limb, the upper part of the Silver Hill has been faulted out, and the thin lower part is covered with slide rock.

Southern part.—Most of the Hasmark in the west limb is simply recrystallized, to as much as 1 mm. at the intrusive contact. Iron is heavily concentrated in a few places near the contact. In the Silver Lode mine,[5] magnetite occurs with limonite, serpentine, and dolomite, in a highly sheared and brecciated area, in part adjacent to the intrusive. The contact of magnetite and marble is sharp and irregular. Stoping indicates large bodies of magnetite along the contact below and above the level of Figure 3. The intrusive rock is rich in hornblende, increasing toward the contact from 25–100 feet away; and at tunnel no. 2 the rock is entirely hornblende. The border zone is locally altered to scapolite or pink clinozoisite.

Another body of magnetite occurs at the Redemption mine, in the Hasmark near its faulted contact with the Silver Hill, about 1000 feet north of the Silver Lode mine (and the intrusive contact). An open cut (no. 1) poorly exposes an irregular mass of intergrown magnetite and ludwigite [$3MgO \cdot B_2O_3 \cdot (Mg,Fe)O \cdot Fe_2O_3$], with serpentinized olivine and chondrodite, in recrystallized dolomite. Two tunnels once cross-cut the magnetite at lower levels, but there was little, if any, ludwigite. Magnetite is also reported in a drift northeast from the Redemption shaft (no. 4); the body of magnetite apparently had good vertical continuity.

At the old Bimetallic Mill, on the east bank of Douglas Creek, a small body of dark augite diorite is veined by hornblendite, identical to that of the Silver Lode mine.

South of Douglas Creek 1000 feet of the granite-Jefferson limestone contact consists of a 60-foot wide band of massive zoned garnet-vesuvianite-calcite tactite. In the limestone, just beyond this band, pits on the Henry Clay claim expose a tabular body of magnetite a few feet thick. In the immediate vicinity the limestone has been recrystallized to cleavage rhombs several inches in width.

[3] Section BB′ is not well exposed on the cliff, but section CC′ correlates it with the excellent exposures of Plate 2,B.

[4] Most of the underground workings shown were inaccessible in 1947–1948; the geology is generalized from maps of the Trout Mining Division, American Machine and Metals Co.

[5] Only tunnel no. 4 was barely accessible; the geology is shown in Figure 3.

On the north side of Camp Creek, just east of the town of Philipsburg, the Red Lion formation shows metamorphism unusual for the western limb of the anticline. Parts of two beds Pl. I) as modified in Figure 2 (rectangle in east-central), the sediments generally parallel the intrusive contact, in marked contrast to the crosscutting near Philipsburg.

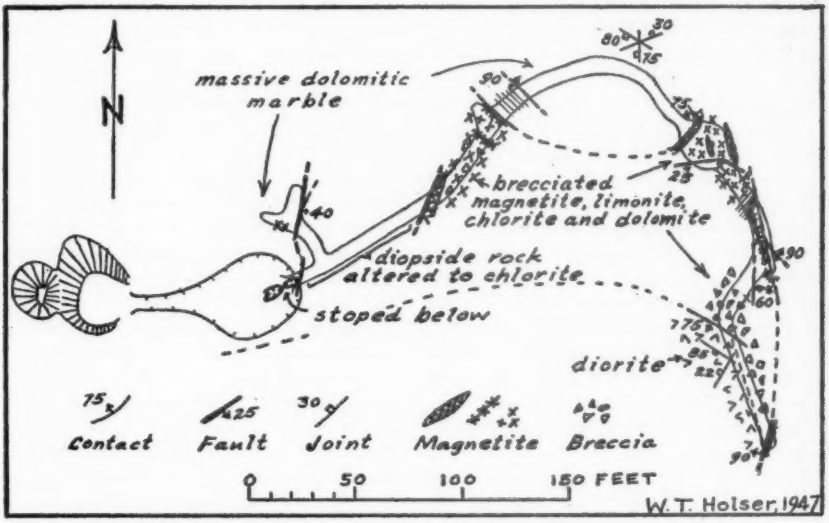

FIGURE 3.—GEOLOGY OF TUNNEL NO. 4, SILVER LODE MINE, PHILIPSBURG AREA
Elevation 6100 feet, western slope Franklin Hill. Note the extensive fracturing and the irregular nature of the magnetite contact.

have been changed to garnet-diopside tactite. Contiguous beds contain diopside, with some tremolite in border areas, and the marble beds are coarsely recrystallized. South of Camp Creek, and also farther up Poorman Hill, most of the Red Lion formation is only slightly recrystallized. Control of this metamorphism was not detected, but its concentration near a sharp bend in the fold suggests control by resultant fracturing. An unexposed intrusion is an alternative explanation.

Just north of the area of Plate 1, on Red Hill, and around the mouth of Stewart Gulch, a poor exposure of the Madison limestone has been metamorphosed to massive garnet tactite.

Other Areas Bordering the Philipsburg Batholith

Except for scapolitization of the Beltian Newland formation at Rumsey Mountain (Emmons and Calkins, 1913, p. 43–44), the southern contact of the batholith shows only slight metamorphism. Referring to Calkins' map (1913,

Just northeast of Twin Peaks, a section of lower Paleozoic is steeply overturned to the southeast. In the bottom of the valley, the lower 75 feet of the Silver Hill is spotted biotite-cordierite hornfels. The upper 150 feet is composed of alternate beds, 3–30 feet thick, of biotite-calcite banded marble and biotite-diopside banded hornfels. All these are identical with rocks found on the western limb of the Philipsburg anticline; but the upper part is only correlative as a whole, and no grossularite marble or tactite occurs. Neither the Silver Hill nor the earlier quartzites outcrop eastward; they presumably are faulted.

The Hasmark dolomite is only about 300 feet thick. Stratigraphically above it, typical Red Lion banded marble contacts the batholith from Twin Peaks to Racetrack Creek (except for a short distance of probable Maywood limestone). The Red Lion shows recrystallization of calcite, and in some places folding of the silica bands, but no silication. The absence of

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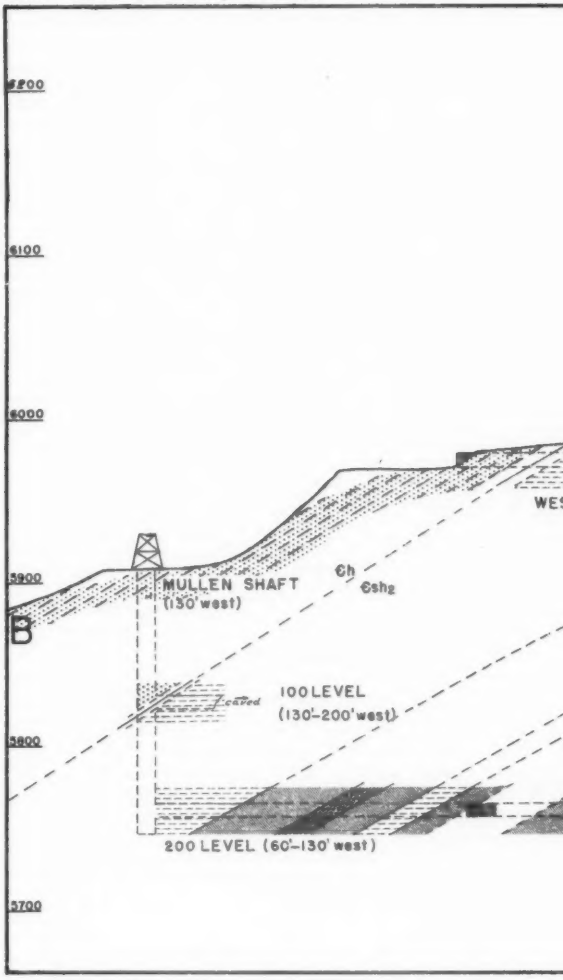
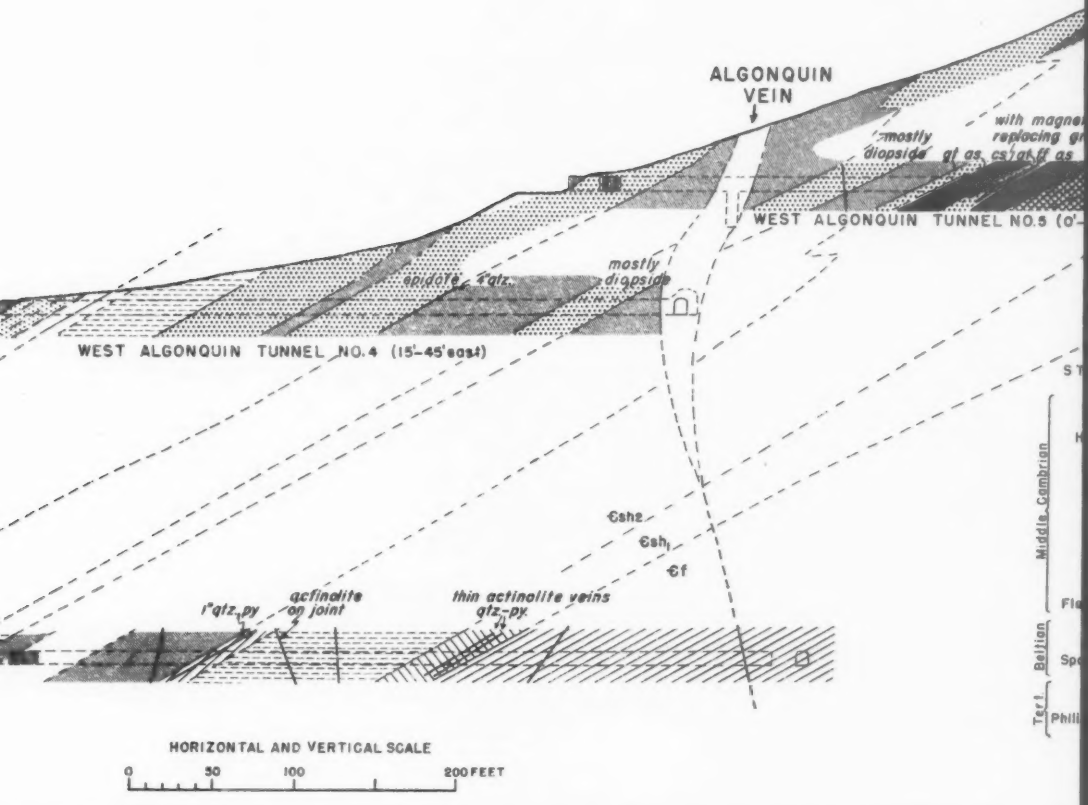


Projected to N. 30° W. Plane



DETAILS OF METAMORPHISM AT WEST AL

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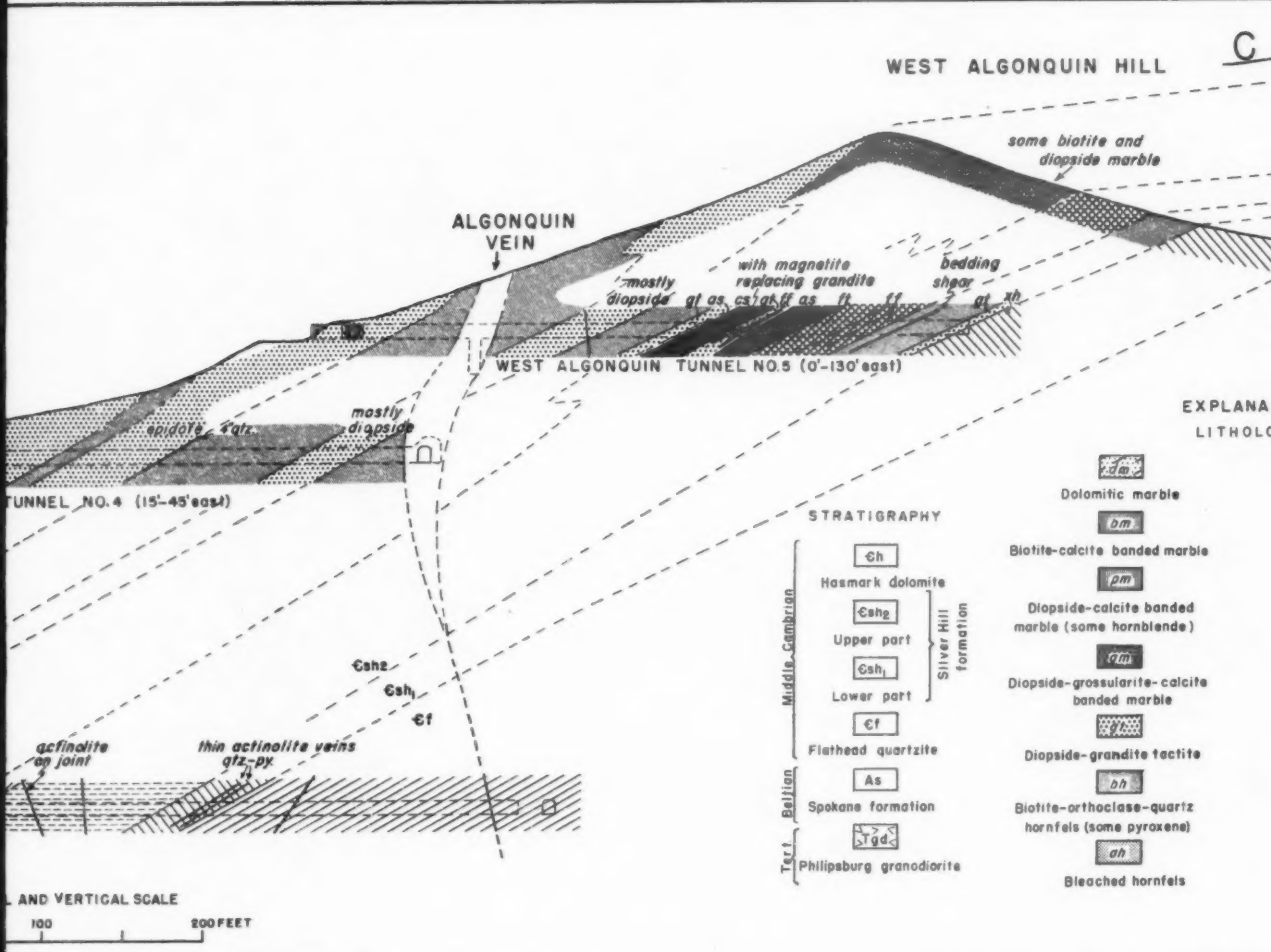
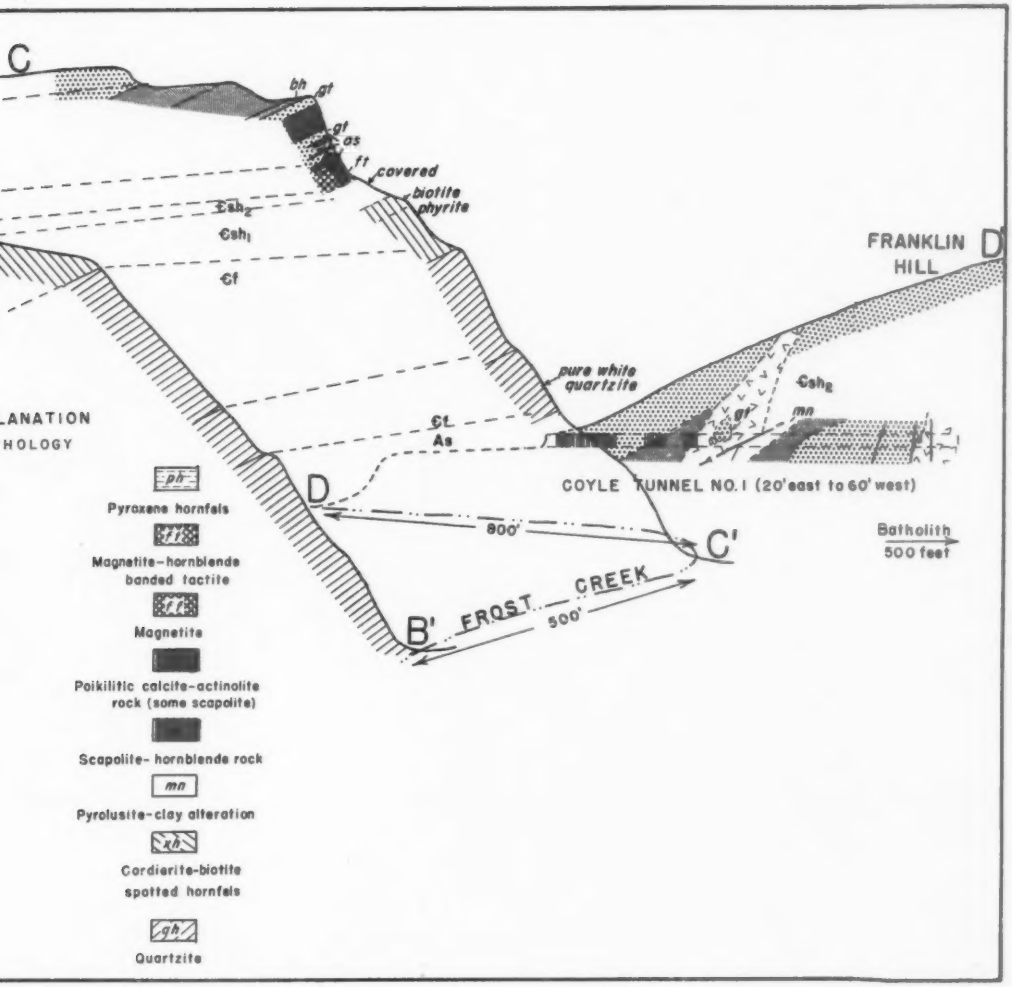


DETAILS OF METAMORPHISM AT WEST ALGONQUIN HILL, PHILIPSBU

Sections Parallel to Axis of Anticline



SBURG AREA

W. T. Holser, 1947

tactite is particularly striking in this formation. Although one can approach within 10 feet of it, the contact is covered, and possibly faulted.

South of Racetrack Creek, the Hasmark is separated by a breccia zone from a section of Madison limestone, which comprises the bulk of the mountain. The latter is irregularly recrystallized, and the characteristic nodules of chert are commonly bordered by a mat of tremolite needles.

South and east of a small lake (Fisher Lake), magnetite bodies occur in the Madison. They are elongate in the bedding direction, and approximately parallel the intrusive contact, which is distant 600–800 feet. They are less than 100 feet long by 20 feet thick. The limestone beds stratigraphically adjacent are apparently identical with the rest of the formation. The ore is magnetite and calcite, with minor siderite and hematite.

Along South Boulder Creek, Jefferson limestone is intensely altered at its contact with a large apophysis of the northern part of the batholith. Coarse grossularite-calcite-magnetite rock, at least 150 feet in width, contains concentrations of magnetite with a N. 35° E. trend, approximately parallel to the contact. Eastward, the Madison limestone is bordered by coarse poikiloblastic scapolite rock.

Southern Cross Cable Area

General statement.—The Cable stock intrudes the same series of Paleozoic formations, and same type of structures, as the Philipsburg batholith (Fig. 2). The intrusive is similarly uniformly granodioritic, with little visible lineation and sharp contacts with sediments. Along the northern edge of the stock, in the Southern Cross-Cable Area, the sediments lie in a southerly pitching anticline (Pl. 14). Fault planes do not crop out, but mine openings are crisscrossed by shear planes, some of which are shown in the detailed maps (Pl. 4; Fig. 4). Movement on these planes is small where evident in exposures; but the absence of beds on the eastern flank of the fold, and thick breccia zones in the Cable mine, indicate larger movements. The intrusion apparently has squeezed the sediments, so that near Cable the Hasmark crops out both east and west of its southerly trend. Near the intrusive on the western flank, all formations are bent sharply around to the north.

At Southern Cross the Red Lion is tactite as far as 1200 feet from the intrusive contact, but this may be attributable to dikes. Tactites in the Trilby tunnel contain unusually large crystals of actinolite, calcite, and magnetite, although magnetite is never found in large bodies in this formation. Even in coarse tactites, bedding is still evident as mineral concentrations. Hasmark crops out as typical fine-grained massive dolomitic marble. In the Southern Cross mine a large amount of (auriferous) hematite and limonite (Emmons and Calkins, 1913, p. 231–234) is probably derived from pyrite. Farther east, near the Oro Fino mine, small irregular bodies of magnetite appear in the dolomitic marble.

Still closer to the contact, at the Pomeroy mine,[6] large bodies of pure magnetite are developed; the magnetic anomaly (*see* Pl. 14) indicates the general distribution of magnetite.

The Hasmark is so massive and recrystallized at this locality that dips and strikes are only approximate and may represent flow banding instead of bedding. The N. 35° E., vertical attitude appears to be a northward exaggeration of the infolded Red Lion and is about parallel to the intrusive contact. The main body of magnetite is tabular and parallels this trend. Magnetite is also concentrated along bands in the dolomite (Pl. 6, fig. 4), which contains much serpentine, probably derived from forsterite. Similar rock probably underlies the poorly exposed ground east to the intrusive contact; apparently the highest concentration of magnetite is not immediately on the contact. Massive diopside-plagioclase rock, probably derived from dolomite, surrounds two small granodiorite dikes in the Pomeroy tunnel. However, it is highly veined and altered with biotite and calcite, making interpretation difficult.

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[6] The Pomeroy and neighboring mines were prospected for magnetite by the U. S. Bureau of Mines in 1944 (Wimmler, 1946). The Pomeroy tunnel was not entered, and the report contains no geologic map. It is unfortunate that only selected cores were saved from the diamond drilling, and that the original magnetometer data are not available. Geology in the open pits, underground workings, and diamond drill holes is shown in Figure 4; the two parts of the figure cover the same horizontal area.

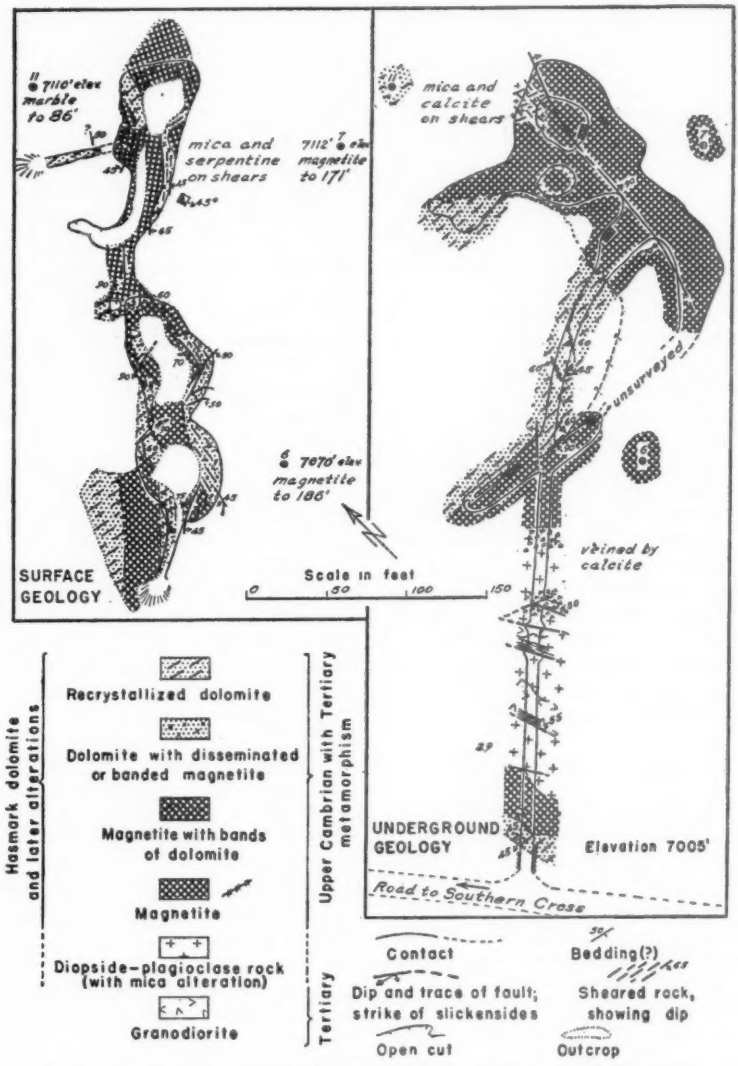


FIGURE 4.—GEOLOGY OF THE POMEROY MINE, SOUTHERN CROSS-CABLE AREA
W. T. Holser, 1947–1948.

Magnetite-dolomite contacts are quite irregular in detail and often grade through rock with less magnetite, as disseminated grains and aggregates. Magnetite and magnetite-dolomite rock contain numerous fractures filled with crystals of a bright-green micaceous mineral, magnetite, and calcite.

In the Silver Hill, near the Duplex mine, the change from biotite marble to tactite is about 300 feet from the contact with the intrusive, which projects into the formation as at Philipsburg. Grandite-diopside and magnetite-hornblende tactite occur, the latter especially in the (stratigraphically) lower part. A magnetite con-

centration apparently extends across the Silver Hill-Hasmark contact (also indicated by the shape of the magnetic anomaly).

The Silver Hill is very poorly exposed on the eastern flank, but float mapping shows the upper member faulted. No tactite occurs in the Silver Hill on this flank.

Cable mine.—The geology of the Cable mine was investigated by Emmons (1906, p. 45–55; Emmons and Calkins, 1913, p. 221–231) when the lower levels of the mine were open. The present study complements that pioneer work with a detailed geologic map (Pl. 4), with particular attention to the occurrences of magnetite and shear zones.

The Cable mine is in Hasmark dolomite, bounded on the southwest by the main Cable stock and on the north and east by a large apophysis (Pl. 14). In the western part of the mine, near the surface, Hasmark dolomite is divided by more granodiorite. The northwestern underground workings are in marble nearly continuously. However, outcrops of Spokane quartzite indicate that marble cannot extend much farther than now exposed on the 6800 level, even if the fault dips northward (reverse). The minimum vertical displacement is 400 feet, downward on the south.

Hasmark exposes little structure in outcrop, but underground extensive faulting is seen, as serpentinized shear planes and zones of coarse breccia. Movement is later than the intrusive, the magnetite, and the recrystallization of the marble (Pl. 8, fig. 3). Breccia zones and many shear planes trend northwestwardly, parallel to intrusive contacts. Breccias are universally present at contacts, and large horses of marble are found in granodiorite. Along the southeastern contact, the granodiorite is severely faulted in a zone about 120 feet wide; farther away it is only well jointed. Downward from the 6800 level marble narrows or becomes more mixed with intrusive; the 6500 level is about half in granodiorite.

The Hasmark is bleached and marmorized to a white or light-gray granular marble, which is still dolomitic. Anhydrous silicates are quite rare. Emmons (Emmons and Calkins, 1913, p. 224–225) attaches some importance to small bodies of garnet found on the 6586 level, but the writer has found none in the accessible levels, or in selected diamond-drill cores of the lower levels. Biotite replaces marble in veins and masses, generally near the intrusive or fault planes; boundaries are sharp but irregular (Pl. 13, fig. 1). Biotite is extensively altered to chlorite, and cut by veins of calcite, especially where sheared (Pl. 8, fig. 2). Both biotite and calcite veins cut magnetite.

The bodies of magnetite are irregularly but distinctly tabular with northwesterly vertical trend, approximately parallel to the contact. Figure 1 of Plate 8 illustrates a strikingly linear vein. The Showers stope mass is vertically continuous between the 6735 and 6945 levels (Pl. 4).

Pyrite, and a little chalcopyrite, pyrrhotite, and arsenopyrite occur in disseminated masses, as in the Lake and Showers stopes, and also in well-defined veins, as on the 6945 level. Native gold has been the ore mineral; its distribution is apparently quite irregular (on the basis of reports and the shape of stoped areas).

On the Atlantic Cable Extension claim (Pl. 14, northeast of the Cable mine), the general direction of strike of Hasmark and Red Lion intersects the intrusive contact, but locally near the contact bedding is deformed. The Hasmark is bleached and recrystallized; the Red Lion is locally garnet-diopside tactite, but generally only marble. Magnetite occurs in the Hasmark as veins a few inches to a few feet thick, forming a line of deposits about 125 feet away from, and parallel, to the intrusive contact. This same line of deposits crosses the Hasmark-Red Lion boundary, and in the Red Lion is found as a zone of magnetite-garnet-hornblende banded tactite.

Daly Gulch Area

The southwestern edge of the Cable stock intrudes the west limb of a northwesterly trending, axially faulted anticline, probably a continuation of the Cable Mountain anticline. Near the contact, however, the Maywood, Red Lion, and Hasmark swing eastward more than 90°, to parallel the intrusive contact (Pl. 6, fig. 2). The Maywood appears both eastward and westward of its trend, appearing to have been "squeezed" into place. A minor thrust fault is further evidence of intrusive force. In Red Lion about 500 feet south of the intrusive, it dips northeastward and strikes northwestward at right angles to the regional thrusts.

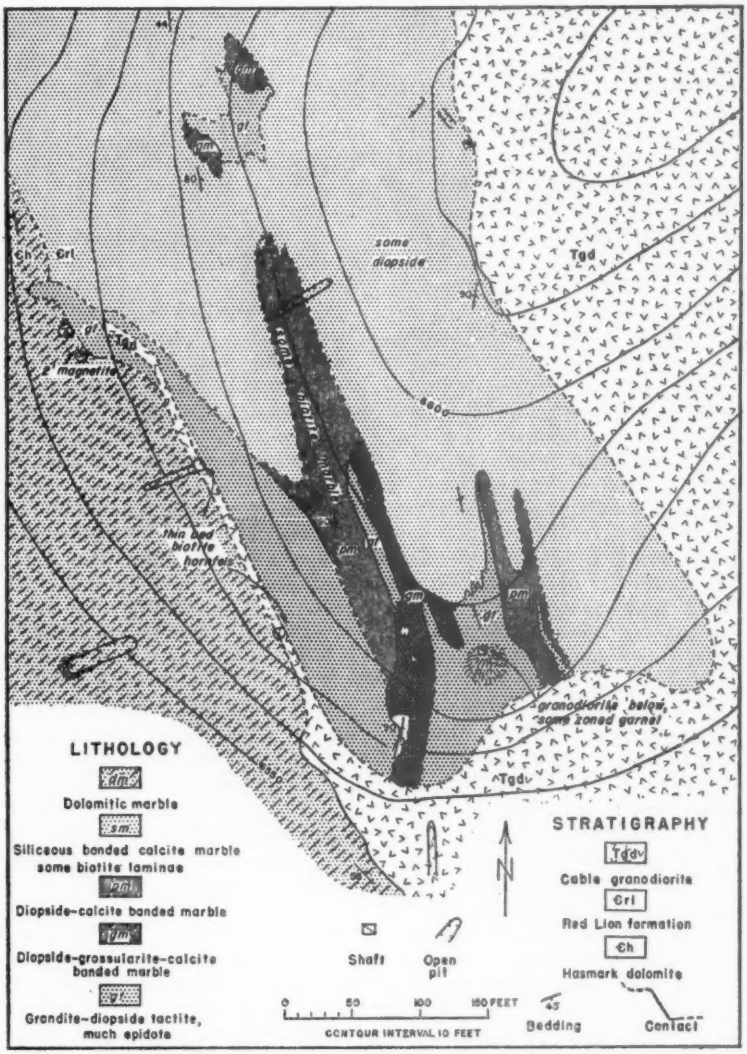
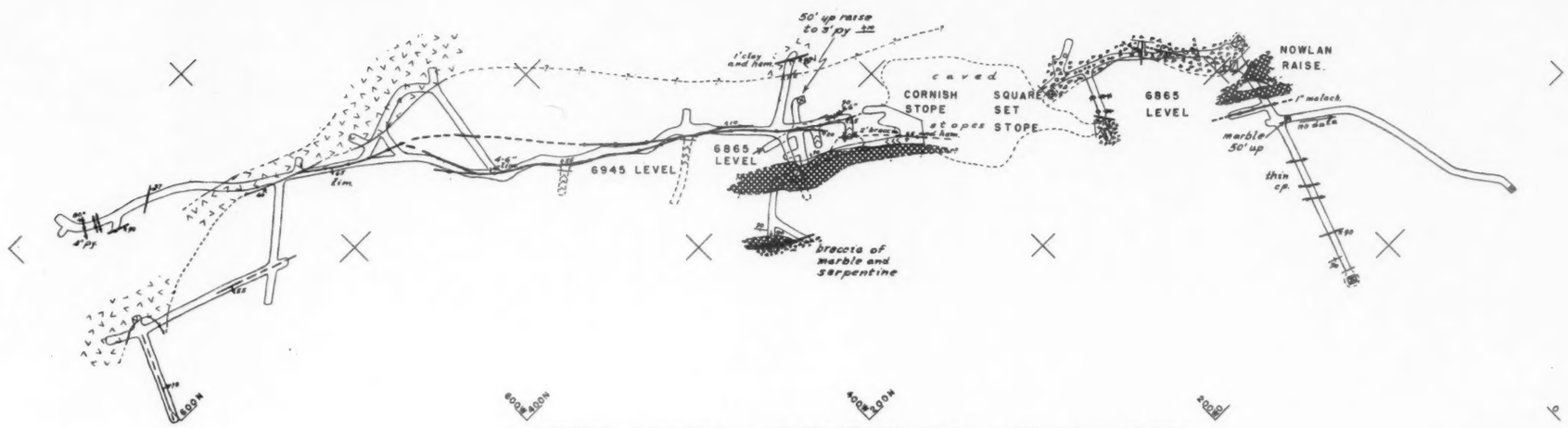


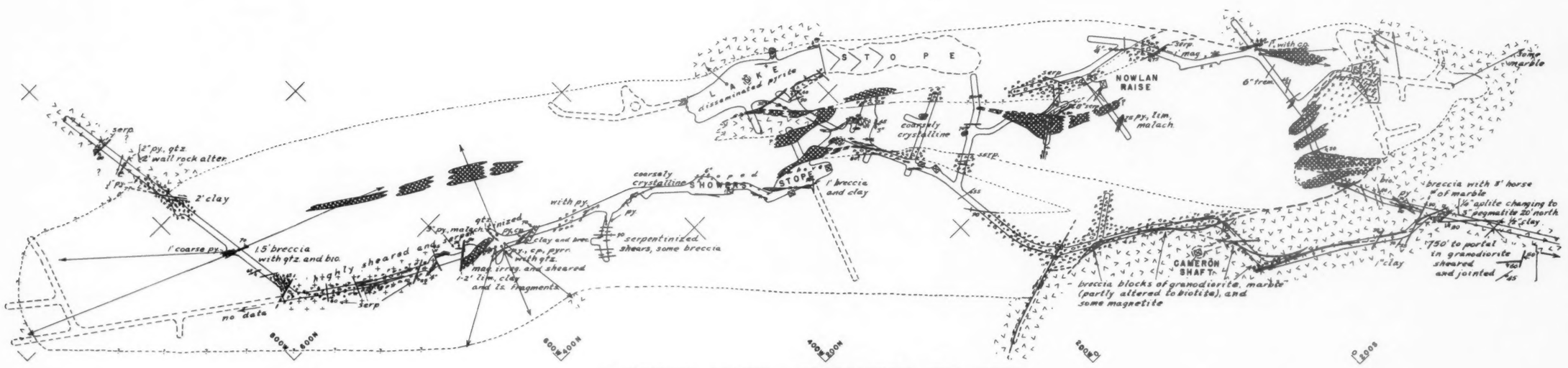
FIGURE 5.—DETAILS OF METAMORPHISM AT CONTACT OF CABLE STOCK, DALY GULCH AREA
J. W. Allingham, 1947; W. T. Holser, 1948.

Metamorphism follows a pattern similar to Philipsburg. The Hasmark shows the usual recrystallization and a small body of magnetite. Red Lion also is merely recrystallized over much of the area, but in some places is a coarse tactite. On the hill at the forks of Daly Gulch (Fig. 5), it is grandite-diopside banded tactite (with much epidote) where the intrusive sharply crosscuts the bedding; where they are parallel, the banded marble is the same as that generally found in the area.

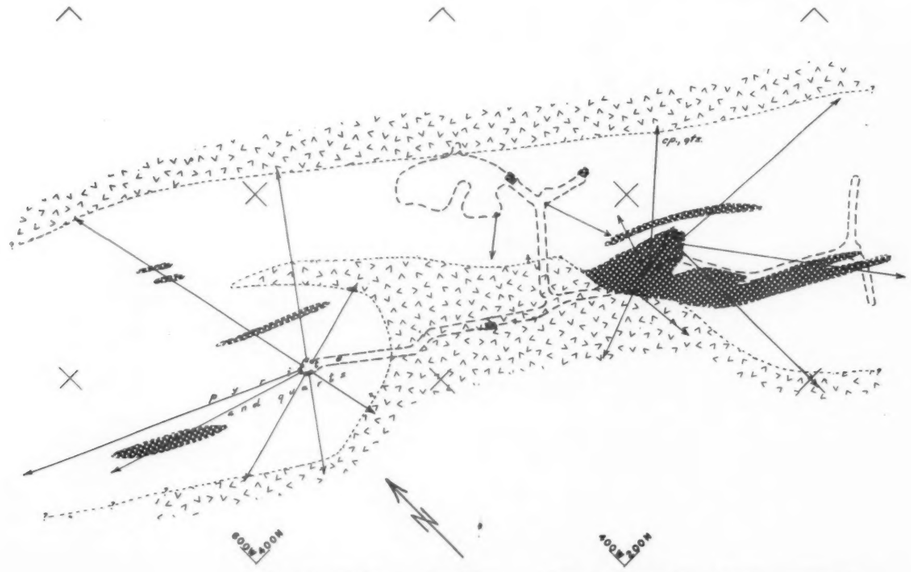
At the contact tactite is commonly of the zoned-garnet type; at another locality half a mile west, similarly situated tactite contains



A UPPER LEVELS - ELEVATION 6865 AND 6945 FEET



B TUNNEL LEVEL - ELEVATION 6800 FEET



C 65-FOOT LEVEL - ELEVATION 6735 FEET

EXPLANATION

- Granodiorite — Tertiary
- Recrystallized dolomite; minor calcite
- Dolomite altered to phlogopite (later alteration to chlorite-serp.)
- Dolomite with magnetite
- Magnetite
- Other vein material
  qtz= quartz; py= pyrite; lim.="limonite"; cp.= chalcopyrite

Hasmark dolomite and later alteration

Upper Cambrian with Tertiary metamorphism

Contact — 65

Joint system

Fault

Sheared rock

Breccia

Drift

Inaccessible workings

Up — Up and down — Down

Raise or winze

Diamond drill hole

0 20 100 200 FEET

GEOLOGY OF THE CABLE MINE, SOUTHERN CROSS - CABLE AREA

Base Map from:
Canyon Lode Mining Co. Maps, 1908, 1917, 1948
W. T. Holser, 1947-1948
Local grid origin near Cameron shaft

Geology by:
W. T. Holser, 1947-1948; J. W. Allingham, Jr., 1947
Cable Lease Mining Co. driller's logs, 1903
Canyon Lode Mining Co. maps, 1917-1948

zoned garnet up to 4 inches in diameter. In the area of Figure 5, the strip of tactite is surrounded by one of grandite-diopside-calcite marble and diopside-calcite marble. The garnet is somewhat higher in andradite content than commonly found in marbles of the region.

Tactite in the northern part of the map area is not related to any apparent intrusion. It spreads out along beds, away from a line trending nearly perpendicular to the bedding. This probably corresponds to a fracture that silication has obliterated.

Olson Gulch Area

In the southeastern corner of the Philipsburg region (Fig. 2), a small granodiorite stock intrudes Madison limestone, with magnetite near the contact. Some of the magnetite occurs in a coarse amphibole-magnetite banded tactite (Pl. 9, fig. 2), in spite of the homogeneous, massive nature of the Madison limestone which it replaces. The tactite is cut by a complex set of shear planes (Pl. 6, fig. 3) along which is concentrated solid iron oxide, now hematite. Much of the magnetite, however, occurs in definite veins that range from a quarter of an inch to 20 feet thick, and appear to replace the limestone along its bedding direction.

Garnet Area

A reconnaissance was made of the contacts of the Top O'Deep stock and the southern edge of the Garnet batholith, because of the similarity of the intrusive and sedimentary rocks to those already investigated, and because of the reputedly large bodies of garnet tactite. The area is 35 miles north of Philipsburg in the Drummond quadrangle; the following discussion refers to the U. S. Geological Survey topographic map of that name and to Pardee's geological maps (1917, Pls. VI, IX).

The situation near the town of Garnet is analogous to that at Philipsburg—a northwestwardly pitching anticline intruded in the northeast by granodiorite. Along the contact to the northwest, in the vicinity of the Nancy Hanks mine, a thin strip of Red Lion formation is metamorphosed to diopside-calcite banded marble, with some garnet-diopside banded tactite. Seven hundred feet from granodiorite, this same formation is the usual recrystallized, but unsilicated, banded siliceous limestone. Much of the float that gave the town of Garnet its name apparently comes from a prominent hill just north. The large mass of garnet-epidote tactite appears at first to be at the intrusive contact with the Hasmark, which outcrops in the vicinity as a coarse white dolomitic marble. However, it is bounded on the south by a thin strip of schist and hornfels; and fault relations in a near-by tunnel of the Shamrock mine indicate that this is a block of Spokane formation; therefore, the garnet tactite is Silver Hill.

Along the western slope of First Chance Gulch, just south of the town of Garnet, the Silver Hill overlies Precambrian hornfels and schist, as a series of cliffs of biotite-calcite banded marble, biotite-diopside banded hornfels, and a more massive recrystallized limestone. Just below the crest of the anticline, about 2000 feet from the intrusion, it appears in a few outcrops as banded garnet-diopside tactite, and continues around to the intrusive contact, just east of the town.

A small granitic stock outcrops a short distance east of the Garnet batholith, in the vicinity of the old placer camp of Top O'Deep. Particularly where the contacts come out on the higher ridges—that is, in the higher parts of the stock—the Madison limestone is locally metamorphosed. Garnet tactite is 2–10 feet wide along some of the intrusive contact. One zone of garnet, less than a foot wide, cuts across the limestone bedding 300 feet from the contact. At the Boston mine, coarse andradite and magnetite are interbanded.

Isochemical Metamorphism[7]

Macrocrystallization[8] *and Chemical Composition*

The distribution of grain size in the pure carbonate beds of the Philipsburg area may in-

[7] Strictly, isochemical thermal metamorphism, absence of directed stress here assumed. Equivalent to "normal contact metamorphism."

[8] Crystal enlargement and crystal regrowth are usually confused or combined. It is suggested that *macrocrystallization* be applied to the former process, and *recrystallization* be confined to the latter. In a recent paper on strained aggregates, Buerger and Washken (1947, p. 304) make no such distinction, but their data indicate that recrystallization was complete at a temperature lower than that at which macrocrystallization began. *Marmorization* is the combination of recrystallization, macrocrystallization, and bleaching, in carbonate rocks.

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dicate some effect of composition. Madison limestone, and to a lesser extent Jefferson limestone, are calcitic, whereas the Hasmark is dolomitic. Allowing for a maximum amount of isomorphous substitution (Winchell, 1933, p. 70), the Hasmark analysis in Table 2 calculates to only 13 mol per cent of dolomite (of total carbonate). Poor solubility in acid and other characteristics indicate that the Hasmark is highly dolomitic, with only local areas of calcite. Fairbairn and Hawkes (1941, p. 620) state that their specimen from Philipsburg "is composed of dolomite crystals only. . . ." My thin sections of Hasmark indicated dolomite, where twinning was present. The specimen analyzed by Schaller (Table 2) probably is not "typical" of the Hasmark.

TABLE 2.—PARTIAL CHEMICAL ANALYSES OF CARBONATE ROCKS, PHILIPSBURG REGION

W. T. Schaller, analyst (Emmons and Calkins, 1913, p. 58, 68)

	1	2
CaO	44.33	50.48
MgO	6.46	0.74
Insoluble	0.45	7.18

1. Hasmark formation, "typical specimen"
2. Madison limestone, black shaly part, Warm Spring Creek

Insofar as measurements can be correlated, Plate 1 indicates that Hasmark has been macrocrystallized farther from the intrusive contact than the calcitic limestones, on the eastern limb of the anticline. No such correlation can be made on the western limb of the anticline, although generally the Hasmark is marmorized to a white, surgary, even-grained rock.

For the Vermont marbles, Bain says:

> "... recrystallization of limestone to calcite marble has not been observed in this marble belt where the magnesium carbonate content exceeds 4 per cent. Beds with more magnesium content ... have remained fine-grained even adjacent to the coarse-textured but purer ... marble." (1934, p. 130.)

An analogous calculation gives only (magnesian) calcite for this marble, so the effect is opposite that at Philipsburg, even if the doubtful analysis is accepted. However, it is not known whether Bain's magnesian limestones are aggregates of calcite and dolomite grains (calcite-dolomite marble) or calcite grains with magnesium in solid solution (magnesian calcite marble). In the case of calcite-dolomite marble the mixture of two phases may inhibit grain growth as it does in some alloy systems. The additional phase presents an energy barrier to the normal process of reduction of bulk free surface energy.

Marmorization in Relation to Deformation

It would be interesting to know how much of the macrocrystallization in the limestones occurred in connection with the folding of the Philipsburg anticline, previous to the intrusion of the granodiorite. Along the eastern limb of the anticline Hasmark, Maywood, and Jefferson are foliated, generally parallel to bedding, but not distinctly enough for field mapping. Fairbairn and Hawkes (1941, p. 620–624) found *c* axes of the Hasmark dolomite with a high degree of orientation perpendicular to the foliation, as well as preferred orientation of $(02\bar{2}1)$ twin planes.

Twinning orientation and macrocrystallization could be related to tectonic forces according to conventional concepts. Presence of twins possibly suggests that there has been little recrystallization since the (pre-intrusive) deformation, presumably responsible for the twin gliding. Two considerations caution against this explanation:

(1) The processes of recrystallization, that follow twinning or other types of deformation, are imperfectly known (Fairbairn, 1949, p. 164). Rinne and Boeke (1908) eliminated twin-gliding planes in deformed calcite and obtained coarser grains, by heating half an hour at 1200°C. in a carbonic acid environment. This temperature is so much higher than expected at a granitic contact that the results have little significance. Even at lower temperatures, however, it seems unlikely that the thermodynamically unstable twinned crystals would be perpetuated, under conditions favoring much macrocrystallization.

(2) Deformation producing twin gliding may be associated with the intrusion, rather than with the earlier folding. Attention is directed to

deformation associated with the Cable stock intrusion and to the partially faulted nature of the granodiorite contact with the eastern limb of the anticline.

In any case, the elastic strain energy of the deformation must have materially aided both recrystallization and macrocrystallization (Buerger and Washken, 1947); and preferred orientation facilitates macrocrystallization, while retaining orientation (Van Arkel, 1936). Thus, the larger grain size on the eastern limb of the anticline could be due to a previous greater degree of orientation by twin-gliding deformation, as well as to elastic strain that was unrelieved by gliding.

In summary, a dilemma is presented by the coincidence of glide twins and macrocrystals. The twins presumably are due to a deformation earlier than the intrusion; the latter should have eliminated them at the same time that it enlarged the crystals. A detailed petrofabric study, not within the province of the present investigation, is necessary to its solution.

Fracture Control of Marmorization

Several small areas of the Hasmark are macrocrystallized to 1.0 mm. on the west side of the Philipsburg anticline, where the usual size is 0.05–0.12 mm. One limited area in the calcic limestone of the Red Lion has crystals up to 4.0 mm. diameter. In these areas the rocks are prominently fractured along northerly trending vertical planes; in fact, Goddard (1940, Pl. 26) shows them as "sheeted zones."

If the fracturing occurred before marmorization, perhaps before the intrusion, and in some way controlled the marmorization, all traces of the earlier fracturing should have been obliterated. The fracturing is therefore later, and the correlation must be explained otherwise. If the present area is susceptible to fracturing because of some external factor, such as its position near the sharply folded part of the anticline, then it is reasonable to assume that it was that susceptible sometime earlier, and furthermore was similarly fractured then. Such earlier fractures localized marmorization, and the marble was later fractured by the same process, helped by the relative weakness of coarser-grained rock.

Both water and the accompanying heat probably were responsible for macrocrystallization around the fractured zones. Solutions moving in bedding planes and fractures have been given major importance in marmorization (Bain, 1934; Hunt, 1924). However, diffusion must be the principal process in the smaller openings (Holser, 1947).

Bleaching of the Madison Limestone

The Madison is ordinarily very dark gray with bituminous matter. On a hill west of Olson Gulch it shows bleaching and marmorization, irregular in its areal control but regular in the relations of marmorization and bleaching. Areas of pure white marble, with 1 mm. grains, are surrounded by a zone of calcite 10 mm. thick, which has been bleached white but is still only 0.1–0.2 mm. in grain diameter. Beyond a sharp boundary, the limestone is the ordinary dark gray with a grain size of about 0.07 mm. The zone of bleaching does not become appreciably thicker as the area of macrocrystallization becomes larger. These relations indicate that bleaching is closely related to macrocrystallization and precedes it by a small margin. The carbon diffuses away faster or at a lower temperature than the atoms of the calcite crystals diffuse to new positions on larger crystals; or perhaps the carbon has to diffuse out before the crystals may enlarge.

In the banded Silver Hill and Red Lion the calcareous laminae are often marmorized without reaction with the shaly or siliceous laminae. Very large cleavages of calcite (up to a foot in width), as found on the Henry Clay claim and at the Cable mine, are closely associated with deposits of magnetite, and because of the known lateness of the magnetite mineralization are referred to a later stage.

Calcium-Deficient Rocks

In fine-grained thinly laminated hornfels of the upper Silver Hill on the Philipsburg anticline, the mineral association is orthoclase-quartz-biotite, the darker bands being richer in biotite and poorer in quartz. (*See* "biotite hornfels" in Plates 1–3.) In the same locality, rocks of the lower Silver Hill contain cordierite, with

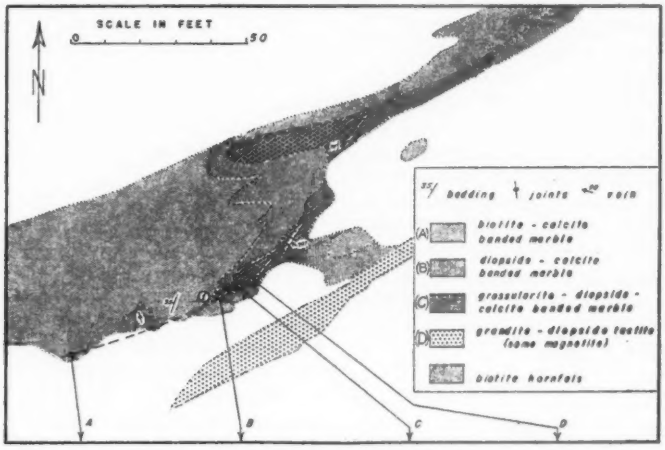


FIGURE 6.—GEOLOGIC RELATIONS
W. T. Holser, 1948. Direction of bedding on the outcrop shown by dashed line.

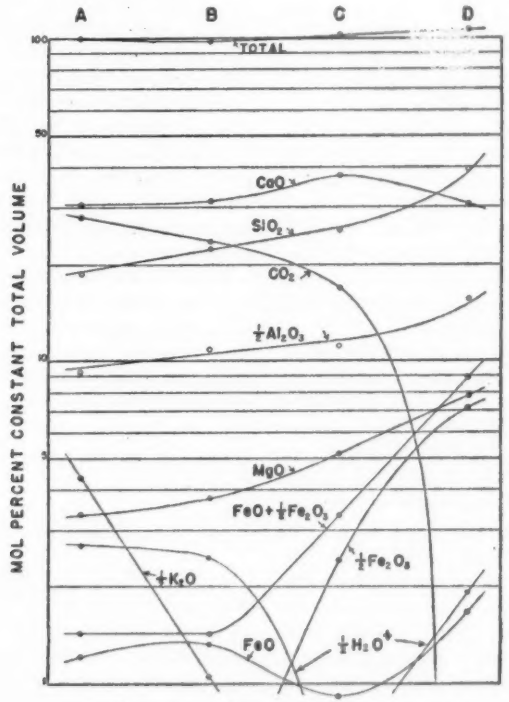


FIGURE 7.—CHEMICAL RELATIONS
(*See* Table 3.)
Details of metamorphism at West Algonquin Hill, Philipsburg area, Silver Hill formation, in vicinity of northeasterly fracture (Point 1 in Pl. 1)

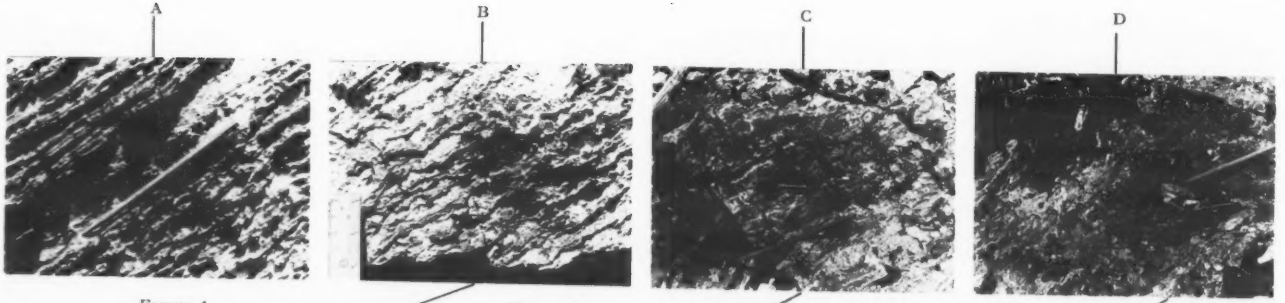


FIGURE 1

FIGURE 2

FIGURE 3

FIGURE 4

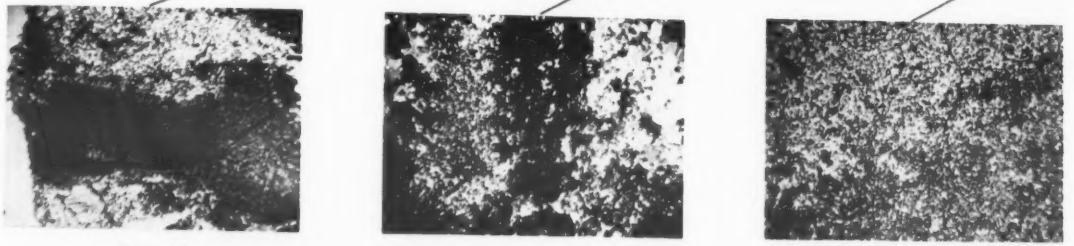


FIGURE 5

FIGURE 6

FIGURE 7

DETAILS OF METAMORPHISM AT WEST ALGONQUIN HILL, PHILIPSBURG AREA: SILVER HILL FORMATION IN VICINITY OF NORTHEASTERLY FRACTURE (POINT "I" IN PLATE 1)

MINERALOGIC RELATIONS

1.— Biotite-calcite banded marble. Biotite-rich bands weathered in relief.

2.— Contact of biotite-calcite and diopside-calcite banded marbles. Lighter-appearing, violet laminae of biotite on left surrounded and replaced by dark-green diopside laminae on right.

3.— Grossularite-diopside-calcite marble, with bedding bands in direction of pencil. Grossularite porphyroblasts line diopsidic laminae.

4.— Contact of grossularite-diopside-calcite marble (lower left), with grandite-diopside tactite, which apparently fills a fissure beyond to the right. Pencil indicates bedding direction, which has been obscured by tactite metamorphism.

5.— Contact of darkly pleochroic biotite on left, with light-colored diopside on right, in lamina of banded marble. Diopside lamina contains some plagioclase. X 4.3 plane-polarized light.

6.— Diopside, grossularite, calcite bands in marble, left to right; bedding plane vertical. Note fine-grained diopside, and admixture with grossularite. X 4.3, crossed polarizers.

7.— Grandite-diopside tactite, at boundary of light-colored, more refringent garnet above, and darker diopside below. Note indefinite boundary at this stage of the metamorphism. X 7.5, plane-polarized light.

knots of biotite, in a quartz matrix. Other varieties of hornfels, particularly at the top of the Silver Hill formation, contain pyroxene (high in diopside) with orthoclase and quartz. Similar rocks are found in the vicinity of all intrusive contacts examined.

While muscovite is present in some of the rocks, aggregation indicates that it is a replacement of cordierite. Where amphiboles are seen, they seem to replace pyroxene. These facts, coupled with the presence of orthoclase instead of microcline, present analogies with the pyroxene hornfels facies. (*Cf.* inner contact zone at Oslo, Sweden; Goldschmidt, 1911, p. 109–110, 154–159.)

In accordance with classical recrystallization theories (Harker, 1939, p. 30), the textural characteristics of the normal hornfelses indicate a lack of sequence of mineral formation. Some rocks which do have such a sequence are also abnormal in their mineralogy and are discussed in connection with tactite metamorphism. The normal hornfelses were apparently derived from their corresponding sediments without appreciable change in chemical composition. Calkins (Emmons and Calkins, 1913, p. 56–57) observed the high-potash mineralogy of the lower Silver Hill hornfels, but his comparison of chemical analyses of metamorphosed and unmetamorphosed samples indicates no change in potash or other constituents. He suspected the presence of glauconite in the original rocks; in the present investigation it was found to compose up to 15 volume per cent of sandstones in various localities north of the Philipsburg batholith.

Because they have been so well described from other regions, the hornfelses were not studied extensively. The correlation of their metamorphic effects with the closely associated marbles may be important, however. Turner (1947, p. 70) thinks the lower limit of temperature for this facies is about 700–750°C., but on experimental evidence Yoder (1950, p. 250) would place it as low as 600°.

Impure Limestones: Isochemical-Reaction Marbles

Relations of several metamorphic types of the impure limestones are strikingly exhibited in some outcrops of upper Silver Hill on the western flank of the Philipsburg anticline. Local geologic relations in the vicinity of point "1" on Plate 1 are shown in Figure 6. Plate 5 illus-

TABLE 3.—CHEMICAL ANALYSES OF METAMORPHOSED IMPURE LIMESTONES FROM THE SILVER HILL FORMATION, WEST ALGONQUIN HILL, PHILIPSBURG AREA

(Ledoux and Company, analysts)

	A	B	C	D
SiO_2	22.08	26.54	28.04	39.56
Al_2O_3	9.33	10.97	10.45	13.41
Fe_2O_3	0.31	0.12	3.65	9.72
FeO	1.54	1.90	1.23	1.93
MgO	2.72	2.96	3.85	5.30
CaO	33.90	34.80	39.07	29.28
Na_2O	0.10	0.09	*	*
K_2O	4.09	0.98	*	*
H_2O^+	0.48	0.44	0.13	0.30
H_2O^-	0.19	0.18	0.06	0.06
TiO_2	n.d.	n.d.	0.40	0.32
P_2O_5	n.d.	n.d.	0.02	0.11
MnO	n.d.	n.d.	0.11	0.22
CO_2	24.37	20.41	13.68	†
Total	99.11§	99.39§	100.69‡	100.21
Apparent specific gravity‖	2.75	2.76	2.96	3.24

A. No. 13.1733. Biotite-calcite banded marble; 2-foot channel sample across bedding at point A (Fig. 6).

B. No. 13.1744. Diopside-calcite banded marble; 2-foot channel sample across bedding at point B (Fig. 6), same horizon as A.

C. No. 13.415. Grossularite-diopside-calcite marble, eastern slope West Algonquin Hill. Mineralogical composition similar to rock at point designated in Figure 6.

D. No. 13.515. Grandite-diopside tactite, eastern slope. Mineralogical composition similar to rock at point designated in Figure 6.

* Alkalies not determined; not present in any other determination. No alkali minerals present.

† Carbon dioxide not determined; not included in water determination. Qualitative tests indicated no CO_2 and no carbonate minerals were present.

‡ Reason for high total not known.

§ Totals 99.63 and 99.91, respectively, if average amounts of TiO_2, P_2O_5, and MnO, from the other analyses, are added.

‖ Including pore space.

trates the various rock types in outcrop and in thin section. Figure 7 charts the variation in chemical composition.[9]

The normal character of the marble, at this locality, is indicated in Figure 1 of Plate 5. This "biotite-calcite banded marble" is a calcic limestone with thin violet-black laminae of biotite-rich material, apparently original shaly limestone. Biotite is commonly in 0.02 mm. flakes oriented parallel to the bedding. It gives the rock its characteristic appearance, where weathering leaves the laminae protruding. Usually other silicate minerals are not present, except for a few small crystals of diopside in the calcareous portion.

At point B in Figure 6, the formation changes abruptly along the bedding; violet-black biotite laminae give way to light-green laminae. They appear first along the boundaries of biotite laminae and are composed of diopside with some labradorite. A large angle of extinction probably indicates alumina in the pyroxene (augite). Where the diopside is in contact with calcite, it is idiomorphic, but boundaries with biotite are highly irregular. Diopside appears to penetrate along boundaries between grains of biotite, and the weight of evidence indicates that diopside replaces biotite. Figure 5 of Plate 5 shows light-colored diopside and dark biotite, constituting laterally related parts of the same lamina.

Figure 7 (A and B) indicates little bulk change of most elements corresponding to this change in mineralogy. The most striking difference is the loss of most of the potash; its removal may have been partly responsible for the change from biotite to diopside. However, diopside is stable in all varieties of the amphibolite facies with excess potash (*see*, for example, Turner, 1948, p. 76–88); and, of course, it is stable in the pyroxene facies. The principal factors causing the reaction have probably been higher temperature and the accompanying greater rate of diffusion, allowing reaction between calcite and biotite laminae.

The more siliceous banded marbles of the Red Lion usually contain tremolite in addition to biotite. Tremolite is also common in sparse, random distribution through the Madison and Jefferson limestones, including some not marmorized. The source of silica and magnesia is not obvious, but presence of large isolated needles of tremolite indicates migration of material from a large volume of surrounding rock. In other cases, the source of silica is apparent, where tremolite mats surround chert nodules, notably south of Fisher Lake. In the Philipsburg area (southwestern part of Pl. 1), chert bodies were unreactive with the surrounding calcite, although very close to the intrusive contact.

Other examples of structural control of isochemical metamorphism might be cited; but for most of the metamorphosed rock, no such control is apparent. Around the Philipsburg anticline, and in the Twin Peaks area, the Silver Hill is nearly all altered to biotite-calcite marble, and biotite-diopside and biotite-quartz hornfels. These rocks seem to be equivalent in conditions of formation to the usual slightly recrystallized, but unsilicated phase of the Red Lion. Their occurrence over a wide area, up to half a mile from the surface intrusive contact, indicates broad-scale control, with nearness to the intrusion predominant.

Intrusion controls the regional pattern of isochemical metamorphism. but fracturing controls it locally. Solutions are undoubtedly of major importance throughout, although they may be water indigenous to the sediment.

Allochemical Silicate Metamorphism[10]

Grossularite Marble

The maps of West Algonquin Hill (Pls. 1, 2, 3; Fig. 6) show large exposures of grossularite marble. The least garnetiferous specimens display the grossularite between the dark diopsidic bands and the white calcite bands as small light-brown dodecahedrons (Pl. 7, fig. 3; Pl. 5). The grossularite was formed at least partly by reaction of the minerals of the two bands. Derivation of this rock from the diopside-calcite marble is evident in Figure 6; similar evidence

[9] Calculated to mol percentage, corrected to constant apparent volume. Volume including pore space is used, as porosity would be expected to change under the action of solutions.

[10] After Eskola (1939, p. 264). Other terms have generally been used, particularly the terms "metasomatic" and "pneumatolytic" with their definite connotations of process.

occurs at numerous other localities on the hill, and at Garnet and Southern Cross. Some of these rocks contain small amounts of vesuvianite.

The grossularite has a refractive index of 1.77 and a specific gravity of 3.50, corresponding to a composition of about $An_{17}Gr_{80}Al_{03}$ (qualitative tests indicate no manganese in any garnets of the region). Changes in alumina and silica in the rock (Fig. 7) are in the same small proportions of the first pair of analyses. The addition of magnesia and the subtraction of carbon dioxide is increased, and lime is added in substantial proportion. The addition of ferric iron (to make a total of 3.3 mol per cent) is the most striking change.

Silicate Tactites

Pure silicate rocks are common in the Silver Hill and Red Lion, and smaller areas are found in the purer limestones. They range from pure diopside to pure garnet rocks, commonly without any other minerals. Rocks higher in garnet are more common, particularly close to intrusive contacts. Garnet in these rocks has a refractive index of about 1.795 and a specific gravity of 3.64, corresponding to about $An_{35}Gr_{60}Al_{05}$. Diopside is faintly pleochroic in pink and green, and makes the rock dark green. The two minerals occur in alternate bands, parallel with original bedding, which is further brought out by much finer grain in the pure diopside bands. Both minerals are coarsely crystalline in places, but the grandite is usually in larger crystals, sometimes up to 2 cm., particularly where the rock has vuggy cavities along the bedding direction (probably due to a loss of volume on decarbonation). Phases transitional to grossularite-diopside-calcite marble indicate, in general, loss of carbonate with growth of garnet. Garnet bands represent original limestone bands.

Replacement nature of these tactites is proved by the mapped relations shown in Figures 5 and 6, where nearly pure garnet occurs in a thin crosscutting strip next to marble. The tactite suite along the eastern flank of the Philipsburg anticline is proportionately much thicker than in the west; and an almost continuous section of tactite occurs immediately adjacent to the intrusive contacts in many formations: in the Daly Gulch area, in the southern part of the Philipsburg area (Pl. 1), and in the Garnet area.

Another origin of grandite tactite is illustrated in Figure 2 of Plate 7. Beds of light-colored biotite-diopside-orthoclase hornfels appear to be replaced by garnet moving out from small (garnet-filled) fractures. Under the microscope a layer of garnet seems to advance along a fine bedding plane in the hornfels. The garnet layer is surrounded by a zone of fine-grained diopside, free of biotite, and contains some diopside. Atop West Algonquin Hill grandite is concentrated beneath a bed of dense, fine-grained biotite hornfels. Small tongues of garnet penetrate the hornfels, especially where the hornfels is jointed (Pl. 7, fig. 4).

Chemical analyses confirm the addition of iron and loss of carbon dioxide, suggested by the change in mineralogy from marble to tactite (Fig. 7). Significantly, iron has been added without the deposition of any scapolite or magnetite.

Formation of the tactite required large amounts of iron, silica, alumina, and magnesia, as well as the removal of carbon dioxide. The smoothness of the variation curves, through the marble, and the geological relations prove that the same solutions, containing the elements listed, were responsible for the formation of both types of rock.

Mobility of several components is also demonstrated by the mineralogical phase rule, as applied to an open system (Korzhinsky, 1936; 1948; Turner, 1948, p. 51–54). In the grandite-diopside tactite, which ostensibly contains six components, just two minerals are developed: only two components are immobile. The entire variation from diopside-calcite marble, through grossularite-diopside-calcite marble, to grandite-diopside tactite is the result of variation in final composition, as produced by the addition and subtraction of elements of varying mobility. Insofar as known, no difference in temperature or pressure is necessary to cause this sequence to develop, although the temperature probably was higher near the source of the altering solutions.

For these same reasons, the usual mineralogical convergence of different rock types is abetted by a chemical convergence. The mineralogy of rocks, of even widely differing orig-

inal composition, converges to a common suite, depending upon the activity of the solutions acting on them. Convergence is well represented in the Philipsburg region. South of Frost Creek, the siliceous limestones of the Red Lion, the shaly and siliceous limestones of the Silver Hill, and part of the Maywood represent wide variations in proportions of components. They have all been metamorphosed to a coarse grandite tactite that contains some diopside. South of Douglas Creek, in the same area, even the highly calcic Jefferson limestone has been metamorphosed to a very similar rock. These are all close to the intrusive. Farther from the contact, some of the incipient processes of convergence may be observed—recall the replacement of hornfels by garnet.

Magnetite Mineralization

Magnetite Tactite

The tactites contain a variable proportion of magnetite, particularly in the lower part of the upper Silver Hill formation, on the eastern part of West Algonquin Hill. (*See* Plate 2B.) Similar rocks are also found in the Silver Hill and Red Lion formations near Cable and Southern Cross (Pl. 14). These rocks are banded in a manner suggestive of the tactite rocks, although when a large amount of magnetite is present the banding is generally less evident (Pl. 9, fig. 1).

Detailed field relations (Pl. 2B) reveal that magnetite has replaced grandite-diopside tactite. Contacts cross the bedding, although the attitude of these contacts is not consistent. Comparison of sections of the Silver Hill, on the two flanks of the anticline, and down the axis, also supports the conclusion. (*See* Plates 1–3.) The stratigraphic position of magnetite tactite on the east flank is approximately equivalent to biotite marbles and hornfelses in the other sections. Magnetite tactite is not so common in the Red Lion as in the Silver Hill but is strikingly developed at one place in the Trilby tunnel, Southern Cross-Cable area.

The pyroxene in magnetite tactite is darker green than commonly found in the grandite tactites and marbles. It is pleochroic with Z = green, X = brownish yellow; and a minimum refringence near 1.71, corresponding to about 60 per cent hedenbergite. Macroscopically, the garnet is usually very dark red to black. The composition is $An_{55}Gr_{25}Al_{20}$, substantially higher in andradite than that in the tactite; it also has a maximum amount of almandite in solution. Hornblende near pargasite (*see* below) is more common than hedenbergite; it has a characteristic pleochroism of blue-green, to green, to yellow-brown.

Microscopic evidence shows the magnetite definitely later than part of the hedenbergite and andradite, for it cuts across broken crystals of the hedenbergite and embays andradite crystals. Magnetite is at least in part later than pargasite, as it often follows cleavage planes of the pargasite (*see* Pl. 9, fig. 4). However, in some places magnetite and pargasite are in interlocking grains that appear contemporaneous. Magnetite apparently replaces grandite along the boundary of a pargasite layer, cutting across single crystals of the grandite (Pl. 9, fig. 3). Associated with pargasite in this section is a humite, probably chondrodite. Humites have been found in various thin sections of grandite tactites; they are usually later than the grandite and fill fractures in it.

Hornblende in Allochemical Metamorphic Rocks

Hornblende is of wide occurrence in the Silver Hill formation, at West Algonquin Hill. In addition to its prominence in magnetite tactite, it occurs in some grandite tactites that are without magnetite (but near magnetite tactite). It also occurs in diopside-calcite marble, especially near the top of the hill, poikiloblastically surrounding diopside crystals in a manner indicative of its replacement origin.

Hornblende occurs in large bodies at the contact of granodiorite and pure marbles, particularly in the vicinity of the Silver Lode mine, in the southern part of the Philipsburg area. A coarse, monomineralic hornblende rock is exposed in the cuts of the mine, at the immediate contact, and the next outcrops to the south expose a granitic rock with at least 50 per cent hornblende. The concentration of hornblende quickly decreases within the body of the intrusion; hornblende normally is not a major constituent of the granodiorite. In one place, several tens of feet south of the contact, a 6-inch "dike" of pure hornblende cuts granodiorite. Hornblende also veins and replaces a

TABLE 4.—CHEMICAL ANALYSIS OF AMPHIBOLE FROM PHILIPSBURG, MONTANA*
(Ledoux and Co., analysts)

	Weight per cent	Atoms per 2400 O atoms		
SiO_2	37.62	568	800	6
Al_2O_3	16.77	298 (232 + 66)		2
			144	1½
Fe_2O_3	6.81	78		
FeO	6.37	80		
MgO	13.72	310		
			396	4
TiO_2	0.34	4		
MnO	0.12	2		
CaO	13.21	213	213	2
Na_2O	3.45	101	124	1
K_2O	1.24	23		
H_2O^+	0.33	33	46	½
F	0.23	13		
P_2O_5	0.11	—		
H_2O^-	0.08	—		
	100.40			
−O	.10			
Total	100.30			

$\alpha = 1.661$ $\gamma = 1.675$
2V = approximately +60°
X = weak yellowish orange (Munsell 10YR8/4)
Y = dusky yellow green (5GY5/4)
Z = light blue green (5BG6/6)
Z > Y > X
$z\Delta C = 22°$

* No. 13.211. Near portal of tunnel no. 4, Silver Lode mine, Philipsburg area.

diorite border phase of the intrusive south of Philipsburg (Pl. 10, fig. 5).

All the hornblendes are remarkably similar in their peculiar blue-green pleochroism, although the absorption varies widely; it is much less in the hornblende of the marble.

A specimen of pure hornblende rock was analyzed, and its components calculated to the normal amphibole molecule (Table 4). This gives the approximate formula $NaCa_2(Mg, Fe'')_4(AlFe''')_{1\frac{1}{2}}AlSi_6O_{23\frac{1}{2}}(OH, F)_{\frac{1}{2}}$.

The nearest species in the classification of Sundius (1946) is pargasite. The analysis corresponds approximately to one of pargasite from Pargas, Finland (Parsons, 1930), but the analysis is unusual in several respects. The silica content is lower, and the soda content higher, than any amphiboles found in limestones, according to the plot by Hallimond (1943, p. 75). The theoretical limit of "vacant spaces" in the lattice is exceeded by sodium-potassium, and the analysis indicates no replacement of calcium by sodium. Most amphiboles with a similar excess are from alkaline igneous rocks. By analogy with the sodic amphiboles, one might ascribe the unusual blue-green pleochroism to the presence of sodium, but an analysis by Sustschinsky (1912) of a hornblende (from tactite) of similar pleochroism shows only 0.28 per cent "alkalis."

The content of water is unusually low, and in combination with the high content of ferric iron indicates alteration toward the "oxyhornblende" type (Barnes, 1930). The complete change is effected at 800°C., but such a high temperature seems unlikely in the geological conditions of this occurrence. In view of the reputed poorness of water and fluorine analyses in general (Hallimond, 1943, p. 70) any inference of temperature based on these figures will be deferred until the analysis is checked. In addition, the green color (however dark) and high angle of extinction are not characteristic of hornblende with a high ratio of ferric to ferrous ions.

On the basis of its geological occurrence and chemical composition, it is believed that the hornblende in the tactites, marbles, and the borders of the granite, was formed by replacement under the action of fluids rich in soda and other components. The formation of a hydrous mineral in place of a previously stable anhydrous one is generally regarded as representing a lowering of temperature (*see*, for example, Bowen and Posnjak, 1931), and such is believed to be the case here. In the Orijarvi, Finland, region, Eskola (1914, p. 228–229) mentions a "hornblende skarn," to which he adduces a similar origin. However, Eskola's mineral differs in several respects, including a notable lack of soda.

The mineral assemblage produced at this stage bears a resemblance to that of the amphibolite facies (Eskola, 1920b, p. 165).

Magnetite-Carbonate Marble and Related Rocks

The large concentrations of iron in the Philipsburg region are deposits of magnetite in more

or less pure marble, rather than magnetite-silicate tactite.[11]

Magnetite ore is usually accompanied by a small amount of forsterite, commonly altered to light-green serpentine. In Figure 6 of Plate 9 (*see* Emmons and Calkins, 1913, pl. XV, B for an illustration of a similar occurrence near Cable), magnetite fills in around olivine grains, in some cases surrounding them. While some magnetite seems to cut across once-continuous olivine grains, it does not follow the same cracks in the olivine that are altered to serpentine.

At the open cut (no. 1) of the Redemption mine, Philipsburg area, chondrodite and ludwigite are associated with the magnetite ore, in macrocrystallized Hasmark dolomite. Chondrodite is in light-brown granular aggregates which in thin section are pleochroic in light yellow. Polysynthetic twinning occurs with an angle of extinction characteristic of the mineral, but more than one humite may be present. Chondrodite replaces forsterite. One area of continuous extinction was apparently a single olivine crystal, but half the "crystal" is now composed of chondrodite, in accordance with a replacement front passing through the crystal. The relation of chondrodite to magnetite could not be ascertained with certainty, but the two are bracketed by the relations:

olivine → *magnetite, chondrodite* → (*shattering*) → *serpentine*

and may be contemporaneous.

Ludwigite is rather remarkable in its restriction to one side of this open cut, despite a diligent search of the entire region. It occurs as radiating clusters a couple of centimeters in diameter, in pure magnetite. In one specimen, however, it forms a straight cross-fiber vein in the magnetite, about 2 cm. wide. It was probably deposited with and following the magnetite. It was not seen in contact with any other mineral.

The magnetite often occurs as coarsely crystalline masses, having rough crystal faces against the recrystallized carbonate. Contacts of magnetite with marble are sharp, compared with those of magnetite tactite. Although some specimens near the contact have a sprinkling of magnetite crystals that may make up as little as 25 per cent of the rock, such zones are quite narrow and represent only a small part of the ore body.

Magnetite with high remanence was observed in a number of outcrops (such as at the Pomeroy mine), but not underground. Although specimens from these localities have not been checked for the presence of maghemite (γ Fe_2O_3), the brownish appearance and mode of occurrence are analogous to those described for other lodestones (Newhouse, 1929), which apparently do contain that mineral. In a unique instance, on the Henry Clay claim (Philipsburg area), the lodestone has the same altered, brownish appearance but is composed of radiating lamellae. The lamellae show a porous structure in polished section (Pl. 9, fig. 5). Magnetite (not lodestone) has been described with lamellar texture from two localities (Geijer, 1936; Schwartz and Ronbeck, 1940, p. 603) and is supposedly pseudomorphic after hematite. Otherwise, crystalline hematite was found at the Stirton mine, but only in small quantities.

Control of Magnetite Deposition, and Relation of Magnetite Tactite to Magnetite-Carbonate Marble

While the texture of the pure magnetite ore boundaries definitely indicates replacement many of the ore bodies are far from irregular. They occur commonly in tabular masses, whose long dimensions have a preferred orientation within a given area. One or more is evident in each of the magnetite areas; the best examples are illustrated from the Cable Mine (Pl. 4; Pl. 8, fig. 1; Pl. 14) and Olson Gulch areas. In these cases the planes of the "veins" parallel the bedding of the surrounding limestone. In other cases, a tabular mass of magnetite seems to follow a fracture. At the Redemption Mine in the Philipsburg area (Pl. 1), it is continuous along the intrusive contact, and also probably continuous for 1600 feet along the Redemption fault, at right angles to the contact. Even in the first-cited examples, the marble is so uni-

[11] As described above, on the Redemption, Kentucky, Silver Lode, and Henry Clay claims in the Philipsburg area; at Fisher Lake in the Twin Peaks-Racetrack Creek area; on the Pomeroy, Cable, and Atlantic Cable Extension claims in the Southern Cross-Cable area; and on the Gold Crown and Bung Your Eye claims in the Olson Gulch area.

form and massive that control was probably by fractures or flow structure, parallel to the bedding, rather than by any physical or chemical differences among beds.

Other areas of magnetite in marble occur less regularly. In some cases, as in the northwest part of the Olson Gulch area, and in the Pomeroy mine, later faulting obscured any regularity the original deposit may have had.

Control of the magnetite tactite phase is shown by its concentration close to fractures, underlying impervious hornfels, in one locality in the Philipsburg area. In the more highly metamorphosed areas, the control of tactite is not apparent. The marked tendency of the magnetite to concentrate along what is apparently one bed of the Silver Hill is evident in the maps and sections (Pls. 1–3). Rising iron-bearing solutions probably replaced the first beds encountered above the relatively irreplaceable quartzites and hornfelses of the Spokane, Flathead, and lower Silver Hill formations.

The tabular bodies of pure magnetite always occur in the pure carbonate rocks of the Hasmark, Jefferson, and Madison. By contrast, all magnetite in the Silver Hill and Red Lion formations occurs as tactite with silicates, usually banded. Neither is any amount of magnetite, beyond the usual early, small crystals, found in the intrusive itself, although it occurs in large bodies in intrusives of other districts (as at Iron Springs, Utah: Leith and Harder, 1908, p. 67).

The relation between the two types of magnetite occurrence is shown strikingly on the Atlantic Cable Extension claim (Pl. 14). The magnetite tactite area of the Red Lion formation is directly in line, and nearly continuous, with tabular magnetite of the Hasmark dolomite. Evidently both are formed around the same fracture zone, by the same solutions, and at the same time. In the western part of the same area, east of the Oro Fino mine, magnetite is found on both sides of the Red Lion-Hasmark formational contact. It is inferred that all (original) magnetite mineralization in a given area is closely related, both in time and circumstances of deposition. This generalization is not intended to apply to redistribution of magnetite in scapolite metamorphism, to be described.

The magnetite phase of metamorphism is probably not separated sharply in time from the preceding silicate metamorphism and the succeeding scapolitization. The microscopic evidence of replacement makes it clear, however, that these rocks were marble and silicate tactite in the process of becoming magnetite tactite. An areal relation such as has been illustrated (Fig. 6) signifies an analogous temporal relation, at a given point.

Chemistry of Magnetite Deposition

Most of the magnetite-bearing carbonate rocks are of the simple magnetite-carbonate-olivine (or serpentine) mineralogy. In a few restricted localities, however, there has been intensive alteration of the marble to amphibole and epidote. Magnesia and silica, as well as iron, have evidently been added, with a resulting composition and mineralogy converging to that of the magnetite tactites.

This phenomenon is well displayed in the northern open cuts in Olson Gulch (Pl. 9, fig. 2). Less than 100 feet from this exposure, the Madison limestone is pure calcite (Table 2). It is also seen, in more limited extent, along the eastern side of the Pomeroy open cuts, in the Cable-Southern Cross area (Pl. 6, fig. 4). These magnetite-silicate tactites are roughly banded in the same direction as bedding planes of the area. In both cases mentioned, the unmetamorphosed rocks are quite massive, and it is difficult to see what might have controlled the banding. The situation is somewhat analogous to metamorphic differentiation (*e.g.*, Eskola, 1939, p. 405–407).

While mineral composition alone should indicate that the formation of magnetite tactites and marbles involved addition of iron, it might be postulated that magnetite was formed by oxidation of ferrous silicates, such as andradite and hedenbergite (note the parallel case in Ramberg, 1948). To prove the addition of iron, an analysis was made of tactite about 30 per cent replaced by magnetite, but containing little pargasite. Comparison of this analysis with the one of grandite-diopside tactite (Analysis D, Table 3; Fig. 7) shows that the total iron content has more than doubled, while magnesia has remained about constant. In the

magnetite-carbonate and magnetite-olivine-carbonate deposits, the addition of iron may be reckoned directly from the magnetite content; no other iron minerals are present. In the tactites of Olson Gulch the addition of magnesia is patent. Magnesia addition was definitely present in the earlier stages of the allochemical metamorphism at Philipsburg (Fig. 7); and Eskola early noted the importance of magnesia in "iron skarn metasomatism" (1920a, p. 106).

TABLE 5.—CHEMICAL ANALYSIS OF MAGNETITE HEDENBERGITE TACTITE, PHILIPSBURG AREA*

(In weight per cent; Ledoux and Co., analysts)

Fe_2O_3	20.43
FeO	15.82
MgO	5.04

* No. 13.439. Lower part of upper part of Silver Hill formation, near top of West Algonquin Hill.

TABLE 6.—CHEMICAL ANALYSIS OF ARTIFICIAL HORNBLENDE

(K. de Kroustchoff, analyst)

	Weight per cent	Atoms	
SiO_2	42.35	640	6½
Al_2O_3	8.11	142	1½
Fe_2O_3	7.91	90	1
FeO	10.11	129	4½
MgO	14.33	325	
CaO	13.21	215	2
Na_2O	2.18	70	1
K_2O	1.87	36	
H_2O*	0.91	92	1
	100.98		

* Loss on ignition.

One of the problems of iron deposits is the determination of the anionic associates of the

PLATE 6.—AREAS OF INTENSE METAMORPHISM

FIGURE 1.—WEST ALGONQUIN CLIFFS, PHILIPSBURG AREA

Second cliff from left edge in Plate 2B. Knapsack at left base of cliff gives scale. Rough-textured grossularite-diopside-calcite banded marble near top of cliff. Near base of tree is white band of scapolite rock, and another occurs few feet lower. Below more marble and tactite is smooth dark surface of magnetite tactite, which on the left passes into coarse white scapolite rock.

FIGURE 2.—SOUTHERN CONTACT OF CABLE STOCK, DALY GULCH AREA

Vertical aerial photograph; width of area 7000 feet; north at top. Abandoned railroad in upper left; Daly Gulch in lower center with area of Figure 5 near "Ꞓh-Ꞓrl". Note normal northeasterly trend of the Hasmark (Ꞓh) and the Red Lion (Ꞓrl) on ridge in lower left, and sharp swing to southeastward, in immediate vicinity of the stock. The Maywood (Ꞓm) swings both northwestward and southeastward along intrusive contact. U. S. Forest Service photograph.

FIGURE 3.—OLSON GULCH MAGNETITE DEPOSIT

Upper pit, showing faulting near the marble (left)—magnetite (right) contact. Magnetite rocks extensively altered to hematite. Block at left is coarse breccia of marble. Knapsack gives scale.

FIGURE 4.—POMEROY MINE, SOUTHERN CROSS-CABLE AREA

Looking northward in northern open pit, width of area about 25 feet. Note banding of magnetite and silicate, running from upper left to lower right, and solid magnetite on left.

PLATE 7.—GARNETIZATION OF IMPURE LIMESTONES

FIGURE 1.—UNMETAMORPHOSED SHALY BANDED LIMESTONE

Red Lion formation, south of Cable Stock and east of Georgetown Lake. Light gray is limestone, gray is shaly limestone, dark gray is coarsely crystalline limestone. Note anastomosing nature of shaly laminae, and manner in which this structure separates the limestone layers into blocks. J. W. Allingham, Jr., photograph.

FIGURE 2.—GARNET REPLACING HORNFELS

Silver Hill formation, West Algonquin Hill, Philipsburg area. Normal bedding parallels the pencil; note garnet spreading out along bedding from thin fractures that cross the bedding.

FIGURE 3.—DIOPSIDE-GROSSULARITE-CALCITE BANDED MARBLE

Silver Hill formation, West Algonquin Hill, Philipsburg area. Grossularite metacrysts occur along border of diopside and calcite laminae, and have almost replaced the latter.

FIGURE 4.—GARNET REPLACING HORNFELS

Biotite-orthoclase hornfels of Silver Hill formation, West Algonquin Hill, Philipsburg area, replaced by dark garnet along thin joints, upward from garnet marble that covers foreground. Pencil gives bedding direction.

FIGURE 2

FIGURE 4

FIGURE 1

FIGURE 3

AREAS OF INTENSE METAMORPHISM

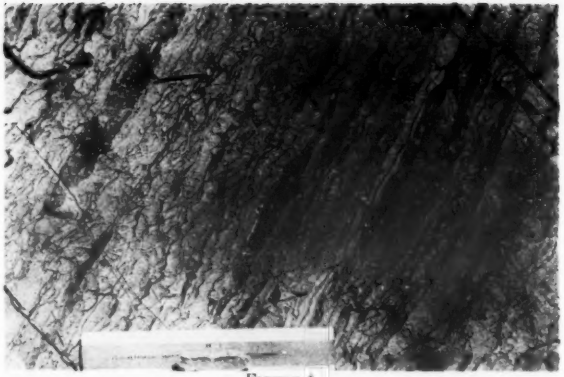


FIGURE 1

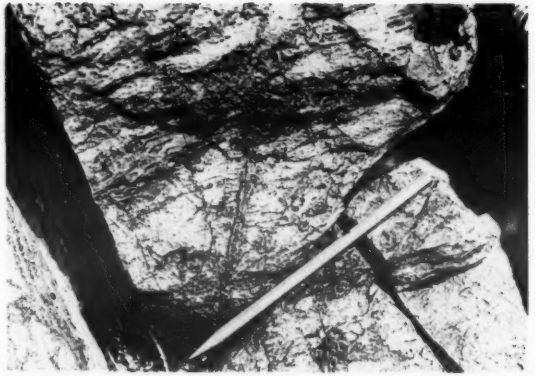


FIGURE 2

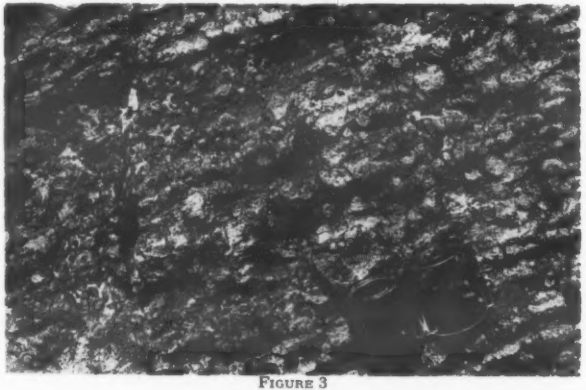


FIGURE 3

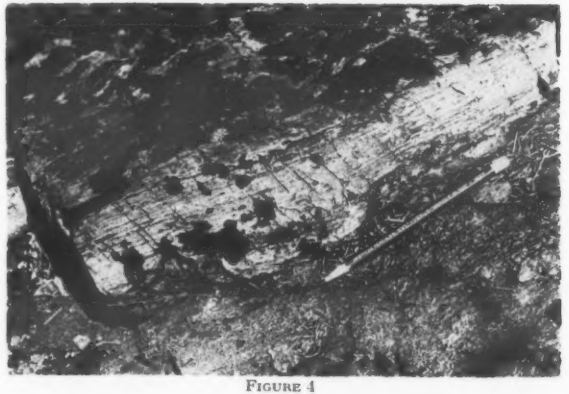


FIGURE 4

GARNETIZATION OF IMPURE LIMESTONE

FIGURE 1

FIGURE 2

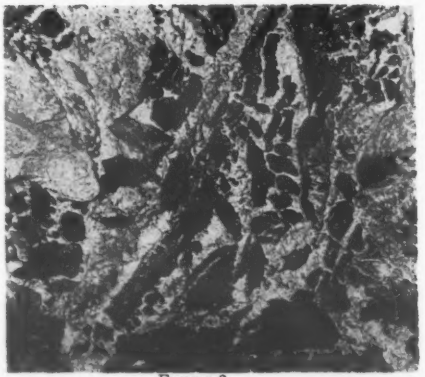


FIGURE 3

FIGURE 4

FIGURE 5

MINERALIZATION IN THE CABLE MINE

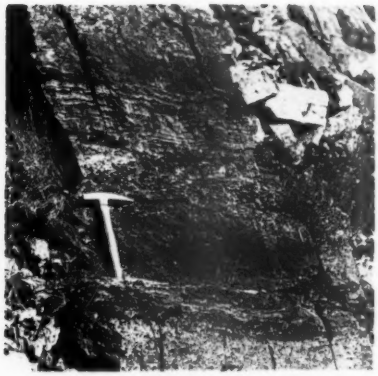


FIGURE 1

FIGURE 2

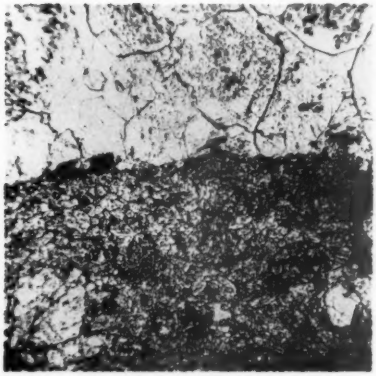


FIGURE 3

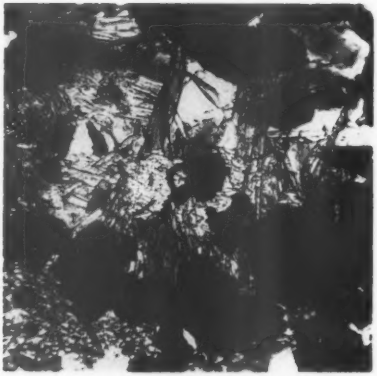


FIGURE 4

FIGURE 5

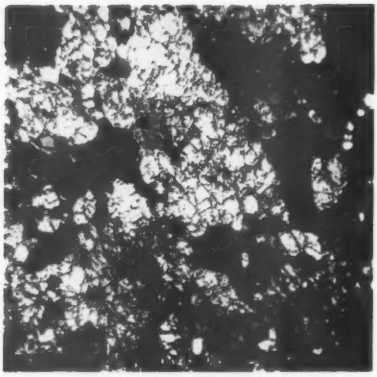


FIGURE 6

MAGNETITE MINERALIZATION

iron, in its fluid transport; some clue may be found in the elements present in the iron ore. The scapolite tactites of West Algonquin Hill are not considered, for reasons apparent later. On the other hand, the closely associated (but partly earlier) pargasite has a fluorine content of 0.23 per cent (mol ratio $F/(OH + F) = 0.31$). Of the 40 hornblendes that have been analyzed for fluorine and water (Winchell, 1945), the frequency mode is at a mol ratio of 0.22. Assuming an isomorphous relation between hydro- and fluor-hornblendes, the solution which resulted in this hornblende must have contained an appreciable proportion of fluorine. Experiments (Grigoriev and Iskull, 1937, p. 170; Barnes, 1930, p. 414) indicate that at high temperatures and one atmosphere pressure fluorine is more tightly bound to the lattice than is hydrogen. Under such conditions, the proportion of fluorine would be less, in a fluid in equilibrium with the hornblende, than it is in the lattice. Synthesis of amphibole from "dry" melts has been successful only with fluorine compounds (Bowen and Posnjak, 1931; Grigoriev, 1935).

Although hydrothermal syntheses of hornblende have been done recently, the only one on which data are available is that of De Kroustchoff (1891) whose product is similar to the Philipsburg pargasite (Table 6).

PLATE 8.—MINERALIZATION IN THE CABLE MINE

FIGURE 1.—MAGNETITE ORE

Back of drift, near main winze, 6800-foot level; carbide lamp gives scale. Note linearity of (northwesterly) trend of magnetite-dolomite marble contact. In detail, however, thin bands of magnetite replace dolomite along parallel and crosscutting fractures. Late calcite veins cut magnetite near lamp. No silicates present.

FIGURE 2.—SHEARING IN BIOTITE-CHLORITE ROCK

Wall of easternmost stope, 6800-foot level; width of area about 3 feet. Marble has been entirely replaced by biotite, which has been chloritized. Numerous shear planes show slickensides in many directions. Late calcite veins on right. J. W. Allingham, Jr., photograph.

FIGURE 3.—BRECCIA

Near Nowlan raise; width of area about 2 feet. Blocks of magnetite and marble, cemented by late calcite.

FIGURE 4.—PYRITE AND MAGNETITE IN MARBLE

Photomicrograph of polished section from occurrence similar to Cable mine, in Hasmark dolomite marble, U. S. Bureau of Mines diamond drill hole No. 2, Hope Hill, Philipsburg area. ×30, vertical illumination. White-appearing pyrite surrounds and veins magnetite. Late chalcopyrite was seen in same section.

FIGURE 5.—GOLD ORE

Showers Stope. ×78, vertical illumination. Pyrrhotite appears gray in lower right. Gold in center surrounded by smoother rim of tetradymite (?). Gold also seen as thin stringer along cleavage plane in calcite at left edge of photograph.

PLATE 9.—MAGNETITE MINERALIZATION

FIGURE 1.—MAGNETITE TACTITE FROM SHALY LIMESTONE

Lower part of upper Silver Hill formation, West Algonquin Hill, Philipsburg area. Gray-appearing silicate layers are dark green, composed of pargasite and hedenbergite, with some grandite.

FIGURE 2.—MAGNETITE TACTITE FROM PURE LIMESTONE

In northern open cut, Olson Gulch area. Dark bands largely magnetite; light (green) bands are amphibole and epidote. In Madison limestone a few feet from contact with Olson Gulch stock.

FIGURE 3.—INCIPIENT FORMATION OF MAGNETITE

Thin section showing magnetite beginning to replace silicate tactite along boundary of grandite (above), and pargasite which has replaced hedenbergite below. Faint peripheral zoning may be seen in grandite. High-index mineral in lower band is humite. ×12, plane polarized light.

FIGURE 4.—MAGNETITE-PARGASITE TACTITE

Thin section of specimen from West Algonquin Hill, Philipsburg area. Note random, and in some cases interpenetrating, texture of pargasite. Small amounts of magnetite may be seen projecting into hornblende, especially along cleavage cracks; in some cases hornblende is nearly surrounded by magnetite. ×34, plane-polarized light.

FIGURE 5.—LAMINATED MAGNETITE

Polished section from Henry Clay claim, Philipsburg area. ×31, vertical illumination. Note cellular texture of magnetite, in comparison with more normal material shown in figure 4 of Plate 8.

FIGURE 6.—OLIVINE-MAGNETITE ROCK

Thin section from Redemption open cut No. 1, Philipsburg area. ×12, crossed polarizers. Black magnetite fills around and possibly replaces olivine, along boundaries of crystals.

The approximate amphibole formula is $NaCa_2(Fe,Mg)_{4\frac{1}{2}}FeAl_{1\frac{1}{2}}Si_{8\frac{1}{2}}O_{23}(OH)$, corresponding to the pargasite type of substitution. In comparison with the Philipsburg pargasite this artificial hornblende contains a similar excess of soda-potash, a similar ratio of Fe″/Mg, and a similar lack of hydroxyl. It also has blue-green pleochroism. It is said that the synthesis took place at 550°C., although Allen, Wright, and Clement (1906, p. 404) insinuate that the temperature was lower. The media were colloidal solutions of SiO_2, $Fe(OH)_2$, and Fe_2O_3; with $Ca(OH)_2$, $Mg(OH)_2$, NaOH, and KOH. The source of alumina was probably the glass vessel. Other phases formed in the experiments were diopside; and analcite (?), quartz, and orthoclase (?) (De Kroustchoff, 1891).

A monoclinic amphibole was possibly synthesized from chloride solutions at 375–475°C. (Allen *et al.*, 1906, p. 405; Bowen and Posnjak, 1931). They also found magnetite, apparently the product of reaction with the steel bomb.

These experiments indicate that amphiboles may be formed in the temperature range 375–500°C., probably within a range of weakly acidic to strongly basic hydrogen ion concentration, and that they are stable in association with magnetite.

Few data are available on the hydrothermal synthesis of magnetite, and this will be one of the principal objects of future laboratory work. The system FeO-H_2O is particularly worthy of investigation (Shand, 1947b); however, there is some doubt concerning Shand's suggestion of hydrothermal transport of iron in the form of ferrous hydroxide hydrosol, because ferrous hydroxide is not stable above 220°C., and in fact seems to have no stability range (Huttig and Moldner, 1931, p. 183–184).

As the magnetite, including that in the marble, was deposited within a short time of the deposition of the pargasite, the magnetite probably was deposited from a fluid concentrated in fluorine. The presence of small amount of phlogopite, chondrodite, and apatite supports this theory. On the other hand, fluorite has never been found in the region, except in minor amounts in a mine on Hope Hill, Philipsburg area.

The distinction is commonly made, in the discussion of allochemical metamorphism at igneous contacts, between "pneumatolytic metamorphism" by a gas phase (usually a volatile compound), and "hydrothermal metamorphism" by liquid phase (usually an ionic solution) (Eskola, 1939, p. 372). However, steam can carry substances in solution, especially above the critical point. Iron undoubtedly can be transported as a halide gas, but it can also be transported in aqueous solutions. It is not known whether such a solution, if it also contained fluorine, would have a sufficiently high vapor pressure of FeF_3 to give an appreciable separation of iron into a gaseous phase. Also, small amounts of halides decrease the stability of Fe_2O_3, in an aqueous system (Sedlatschek, 1942). Finally, it is difficult to see what the differences in effect would be of deposition from gaseous, liquid, or supercritical solutions, that would serve as criteria to differentiate the presence of these states. In view of these considerations, the mode of transport may be characterized as a fluor-aqueous solution, containing a high concentration of iron, with silica, magnesia, and alumina; weakly alkaline hydrogen ion concentration; and state undetermined.

In a carbon dioxide environment, as near the contact of the pure limestone, magnetite is the stable phase, especially below 700°C. (Hawley and Robinson, 1948, p. 608). However, with "a very open system and unlimited quantities of CO_2" hematite might be formed. If such conditions occurred locally and early, perhaps as a result of faulting at the time of the intrusion, hematite would be deposited, to be converted to magnetite under later, more normal conditions. This might result in the bladed form of magnetite described above. The temperature variation of equilibrium constant is such that the occurrence would be more likely at higher temperatures. Occurrences in other regions indicate that, under certain conditions, formation of crystalline hematite in marble at granitic contacts may be widespread.

Late Magnetite Veins

Space-filling veins were observed in the magnetite areas, especially in the Pomeroy and Cable mines, Southern Cross-Cable area. In many cases the veins appear not to have been filled completely with magnetite, and the space has been taken up by later calcite and a green

micaceous mineral; in other cases it is empty (Pl. 13, fig. 2).

The as yet unidentified micaceous mineral varies from greenish white to grayish green, colorless to pale brown in thin section. It has soft, inelastic cleavage flakes. Its composition is near antigorite, but with much FeO, and the refringence is very low (N_m, N_g = 1.530–1.540); the optic angle is about −5°. The mineral is undoubtedly hydrothermal, and later than the serpentine. Emmons (1906, p. 48) called it "green biotite."

Some late magnetite is sensibly anisotropic under polarized vertical illumination, a condition not yet adequately explained. (Strauss, 1946.)

These fissure veins are undoubtedly younger than the replacement bodies. They were not observed in relation to other phases of alteration, so their position is not certain in the following sequence.

Aplite Sills

Occurrence

Aplites are scarce in both the intrusive and surrounding sediments; in the Philipsburg area they are absent from most formations. The Silver Hill is a marked exception for, in addition to the large dikelike apophyses of granodiorite (Pl. 1), a large number of small aplite intrusions occur in the eastern limb of the anticline (Pl. 2). They are nearly all sills, with some feederlike dikes. They vary from a fraction of an inch to 1–2 feet thick, and their boundaries with the surrounding sediments are generally quite irregular. In the main batholith of the Philipsburg area aplite is much less abundant, except for a few "marginal fissure" fillings (Balk, 1937, p. 101). In the Cable mine, the main dikes (Pl. 14) are not aplitic and are probably more closely related to the main intrusive action.

Composition

The aplites vary somewhat in composition, but usually they are light gray, with a sprinkling of feldspar, hornblende, and a few biotite phenocrysts. The feldspar phenocrysts are zoned, increasingly calcic outward from an oligoclase core. The fine-grained groundmass is anhedral. Small hexagons of quartz were seen in one section. Much alteration has occurred, particularly of orthoclase to sericite.

The aplite sills are not a likely cause of the metamorphism to silicate and magnetite tactite, with which they are closely associated on West Algonquin Hill. Metamorphism of sediment blocks, included and rotated in the aplite (Pl. 10, fig. 1), is not detectably different from that shown at a distance from an intrusive. Metamorphism in inclusions is uniform, except for a one-fourth inch margin of grandite, in some specimens. Furthermore, the aplites show no evidence of the mineralization that increased the iron content in tactite, and eventually formed magnetite tactite. In the one locality where magnetite tactite and aplite are juxtaposed, the aplite cuts across many magnetite laminae. On the other hand, aplite is never found with walls of marble.

The immediate contact of aplite with tactite varies from highly irregular to labyrinthian (Pl. 10, fig. 3). The texture strongly suggests intrusion by replacement. A border of garnet near the contact, a fraction of an inch thick, is characteristic. In the "contact zone" on the aplite side of the garnet rim, remnants of diopside and garnet crystals have been highly embayed by the minerals of the aplite, here nearly entirely feldspar. Some of the larger diopside crystals are only altered to hornblende.

Another aspect of the intrusive process is illustrated in figures 1 and 2 of Plate 10. In the second figure, small apophyses of the sill have apparently "pried up" pieces of tactite bed. In the first, two pieces of tactite nearly a square foot in cross section are surrounded by aplite, and the laminae are at a large angle to the normal attitude of bedding in the surrounding tactite. These pieces have been separated from the wall rock, although it is improbable that they were suspended in the aplite.

Aplite apparently intruded by a dual process of reaction-assimilation and mechanical stoping. In tiny pilot dikes near the reaction front, the rock may be solid, or at least a semisolid mesh of partly replaced crystals of wall rock. The pilot sills gradually enlarge by reaction and assimilation, which may continue until the stoped piece is consumed or cooling stops the process. If such is the case, it is not a process of overall replacement, as the observed inclusions were of relatively constant composition, though

of varying size. Goodspeed (1948) has described dikes of similar complex origin.

The replacement aspect of the intrusion would suggest little dilation in connection with its action, and in the few observable cases inclined beds were not displaced.

Endomorphism of the Intrusives

Rims of clear oligoclase on inclusion-rich orthoclase may be evidence of addition of calcium and sodium to the aplite. These are few, and highly calcic feldspars are not found in the sills. Therefore, the continuation of intrusion by this assimilation process would necessitate the continual addition and removal of elements, from and to the dike's source, in order to maintain the alkalic composition of the aplite.

Replacement of hornblende and biotite by pyroxene has been observed both in the aplite of West Algonquin Hill and in the border of the main intrusion to the south. Figure 4 of Plate 10 illustrates a remnant of biotite in the intrusive at the Silver Lode mine. The biotite crystal is rimmed by pyroxene, the inner parts of which are semicontinuous in their extinction, while some of the fragments farther out have other orientations. Hornblende, which the pyroxene also replaces, is the green-brown type characteristic of igneous rocks.

Near Douglas Creek, a small area of intrusive rock is composed of subhedral grains of augite in a groundmass of andesine. It is cut by veins of the blue-green pargasite, and around the veins has been replaced by the same mineral. At the boundary of augite phenocrysts, the hornblende has formed a reaction rim of lighter-colored diopside (Pl. 10, fig. 5).

Pyroxene is not a normal mineral of the Philipsburg batholith. These occurrences of pyroxene seem to indicate a local, late surge of high-temperature solutions, but evidence of this is not found in other rocks. An alternate explanation would have the alteration to pyroxene during the main high-temperature phase of the metamorphic action, after the intrusion. Its parallel occurrence in the aplite would then necessitate an early age for that rock, which does not show any alteration by the late, iron-bearing fluids.

Zoned Garnet Tactite

General Features

Large crystals of garnet with peripheral zoning characterize a rock that occurs at intrusive contacts, with considerable epidote and vesuvianite.

It is more common in the impure limestones of the Red Lion and Silver Hill formations; the most notable occurrence is in Daly Gulch, west of the locality of Figure 5. Figure 1 of Plate 12 shows a light-colored mass of the zoned-garnet tactite, cutting across the bedding laminae of the dark grandite-diopside tactite (Red Lion). Zoned-garnet tactite occurs in the Red Lion formation south of Frost Creek, in the Jefferson formation south of Douglas Creek, Philipsburg area; and a small amount in the Madison formation at Olson Gulch. All these occurrences are within a few tens of feet of intrusive rock. A little occurs at the contact of some of the aplite sills with the Silver Hill, in the Philipsburg area.

Mineralogy

Dark-red to reddish-black garnet is in dodecahedrons up to 3 inches in diameter; faces are best developed where cavities are now present. Refractive index is a little higher than 1.87, corresponding to a composition of about $An_{80}Gr_{05}Al_{15}$, in marked contrast to garnet of silicate or magnetite tactites. Even in the same hand specimen, garnets of the normal massive and zoned types vary as widely in composition.

Under the microscope the andradite shows twinning of the "dodecahedral type" (Goldschmidt, 1911, p. 380; Winchell, 1933, p. 181), composed of 12 rhombic-based pyramids of differing optical orientation. In addition, it exhibits very striking, narrow concentric zones of varying birefringence, some zones being isotropic (Pl. 12, fig. 3). The anisotropic zones are extinct parallel to zoning and to the corresponding surface form. The zones are also visible with plane-polarized light, because of varying refringence. The core seems to have a higher refringence, but fragment measurements revealed no consistent variation in refringence. In many occurrences, the zones are more or less continuous along a broken plane, rather than entirely around individual crystals (Pl. 12,

fig. 2). This suggests a plane (fissure) source of the solutions responsible for them, if not actual growth into such a fissure.

On many outcrops the concentric zoning of garnet crystals is emphasized by a color difference (Pl. 12, fig. 2). The core is commonly black, occasionally dark red or green. In nearly all cases the rim is dark reddish brown, especially where it forms faces in cavities. The intermediate zones vary from red to white.

Under the microscope, the white intermediate zones are seen to be composed of feldspar, calcite, or sometimes epidote; the initial impression is of successive overgrowths. Although in some cases a large part of a zone may be composed of a single crystal of calcite, the texture indicates that the calcite, feldspar, and epidote have replaced garnet in forming the intermediate zones (Pl. 12, figs. 3, 5). Another replacement mineral is isotropic, surrounding the calcite in one of the anisotropic zones (Pl. 12, fig. 4). It has a lower refractive index than the garnet and is probably a hydrogarnet (Yoder, 1950, p. 242).

Lateness of the epidotization is proved by bodies of epidote rock which crosscut single zoned-garnet crystals. In other cases separate garnet crystals have zones replaced by epidote; some crystals are so much replaced that they are recognized only as rounded green masses of epidote. Replacement of garnet by epidote and other minerals, beginning with certain zones of the crystal, may eventually consume the entire crystal.

While the outside boundaries of the replacement zones are usually sharp and straight, along the direction of the dodecahedral faces, the inside boundaries are irregular, with many indentations and embayments into the garnet of the next zone or the core. Some plagioclase also fills in between the outside faces of the garnet crystals.

Anisotropic zoned structure in garnet has received little attention since the inconclusive work of Goldschmidt (1911, p. 379–385; *see also* Ingerson and Barksdale, 1943). The original anisotropic zoning is probably due to intermittent presence of a trace impurity (Frondel *et al.*, 1942, p. 728) in solutions which formed the zoned-garnet tactite by open-space filling or possibly by replacement of garnet-diopside tactite. The presence of impurities is also indicated by the twinned relation of the parts of an abnormally anisotropic zone, as similar twins have been produced in artificial salts (Bunn, 1933, p. 571). Spectrographic analyses are now being made of zones from a single garnet crystal. If the twins are due to surface adsorption of impurities, as deduced by Bunn, they must have been crystallized quickly, and directly from solution.

The garnet-crystal zones composed of other minerals have not been described previously. Their origin by replacement of certain zones of the garnet crystal is evident. The replaced zones are poorly correlated with isotropic zones; the reason for this selection is unknown. The final movement of replacing and replaced material was by solid diffusion, probably along parting planes between dedecahedral twins. Alteration of andradite to feldspar and calcite involved removal of iron and addition of alkalies.

Scapolitization

Areal Distribution

Rocks containing scapolite have wide distribution along the southwest side of the Philipsburg batholith, and in minor amounts in the Southern Cross-Cable area and other localities. They are best exposed on West Algonquin Hill, particularly on the eastern limb of the anticline, as already described (Pls. 2, 3).

In the closely associated poikiloblastic calcite rock, the large calcite crystals enclose a multitude of randomly oriented needles of actinolite, and lesser amounts of scapolite. This rock is also found near contacts south of Douglas Creek, in the town of Garnet, and at the Pearl mine (Top O'Deep district). A coarse magnetite-calcite-actinolite marble from the Trilby mine is probably an allied type.

Structural Control

On the western limb of the Philipsburg anticline, scapolite is concentrated in certain thin beds where they butt against one edge of the outcrop, which exposes a joint plane with clusters of scapolite prisms (Pl. 11, fig. 2). The scapolitization clearly involves replacement of the biotite-quartz-orthoclase hornfels by elements from solutions in the joint plane, and precipitation in the open joint.

Along the eastern limb of the anticline, large-scale mapping has established that fractures controlled the scapolitization (Pl. 2B). Scapolite rock cuts across the bedding, especially toward the top of the hill. The extreme irregularity of the scapolite boundaries on Plate 2B is due to the plane of the cliffs being in about the same direction as the boundaries (Pl. 3, section CC′). The average attitude of the controlling fractures is easterly, with a steep northerly dip. However, the scapolite bodies show wide divergence in their attitude; and they are crosscut by subsidiary flat-lying fractures, which also carry scapolite. Some of these crop out in the cliff next to the left in Plate 2B. (*See also* Pl. 11, fig. 1.) The fractures are filled with crystals of scapolite, and they continue diagonally into the unscapolitized tactite, where they contain epidote as well as a little scapolite (Pl. 11, fig. 3). The garnet surrounding these subsidiary fractures is heavily epidotized, suggesting that epidotization precedes scapolitization.

Paragenesis

Where they are found opposite rocks of such varying composition as grandite-diopside tactite, magnetite-hornblende tactite, and even grossularite-diopside-calcite marble, the crosscutting scapolite rocks are similar in composi-

PLATE 10.—INTRUSIVE ROCKS

FIGURE 1.—INCLUSIONS IN APLITE

Upper right outcrop in Plate 2B. Grandite and diopside bedding bands of wall rock parallel pocketknife in lower left corner. Note size of inclusions, and rotation of their bedding with respect to that of wall rock. White aplite continues as sill to left and right of photograph.

FIGURE 2.—STOPING ACTION OF APLITE

Small fingers of white aplite have advanced along bedding planes of grandite-diopside tactite, with an irregular texture suggestive of replacement. Yet the pieces of tactite have been bent into the sill by this action.

FIGURE 3.—REPLACEMENT TEXTURE IN APLITE

Near crest of West Algonquin Hill, Philipsburg area. Light-colored aplite penetrating magnetite and grandite-diopside tactites parallel to their bedding. Note extreme irregularity of contacts.

FIGURE 4.—BORDER ZONE OF GRANODIORITE

Thin section from Philipsburg batholith at Silver Lode mine, Philipsburg area. ×26, plane-polarized light. Dark-gray, highly pleochroic phenocrysts of biotite replaced by rims of light-gray (green) augite, in groundmass of quartz and oligoclase rimmed by albite.

FIGURE 5.—PARGASITE VEIN IN DIORITE

Thin section from border of Philipsburg batholith, Douglas Creek, Philipsburg area. ×12, plane-polarized light. Hornblende (hb) is replacement vein in dioritic border phase of intrusive. The rock is an intergrown mass of augite (a) and andesine (p). At edge of vein augite has been altered to lighter-green diopside.

PLATE 11.—SCAPOLITIZATION

FIGURE 1.—SCAPOLITE REPLACING MAGNETITE TACTITE

Outcrop second from left, in Plate 2B. White scapolite rock irregularly replacing magnetite tactite along bedding direction. Near pick, a thick scapolite bed passes into magnetite rock with a little scapolite and epidote, on left.

FIGURE 2.—SCAPOLITE REPLACING HORNFELS

Western end of West Algonquin Hill. Scapolite rock weathers dark gray and rough, and may be observed projecting into biotite hornfels, along direction of bedding (shown by pencil point). Straight edge of outcrop is edge of joint plane, covered with coarse scapolite.

FIGURE 3.—SCAPOLITE REPLACING SILICATE TACTITE

White scapolite rock replacing grandite-diopside tactite, which here has been epidotized. Thin, flat-lying fracture filled with scapolite in scapolite rock, and with epidote and scapolite in epidotized grandite rock. Rubber eraser gives scale.

FIGURE 4.—SCAPOLITE ROCK

Radiating white prisms of the mineral constitute the entire rock.

FIGURE 5.—MAGNETITE-SCAPOLITE ROCK

Pod of coarse magnetite rock, with few prisms of scapolite. Such concentrations of magnetite occur only with scapolite.

FIGURE 6.—SCAPOLITE REPLACING MAGNETITE-PARGASITE TACTITE

Thin section in plane-polarized light, ×34. Highly resorbed crystals of pargasite in poikiloblastic mass of colorless scapolite. Where pargasite is armored by magnetite, it has not been replaced. Magnetite only slightly replaced by scapolite.

FIGURE 1

FIGURE 2

FIGURE 3

FIGURE 4

FIGURE 5

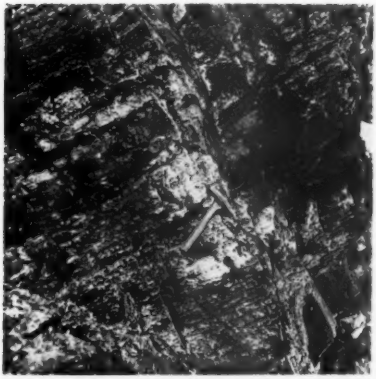
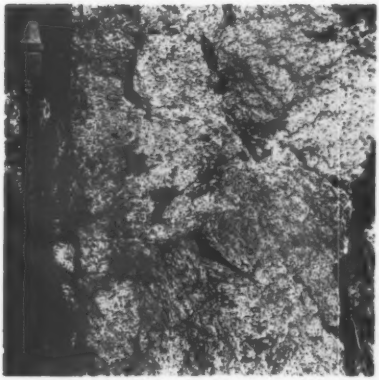


FIGURE 1

FIGURE 2

FIGURE 3



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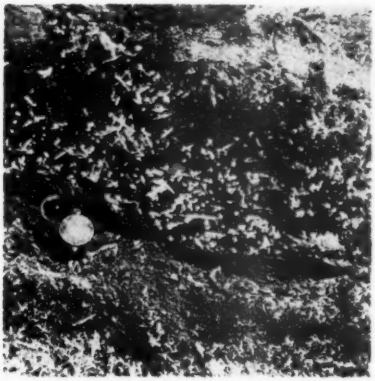

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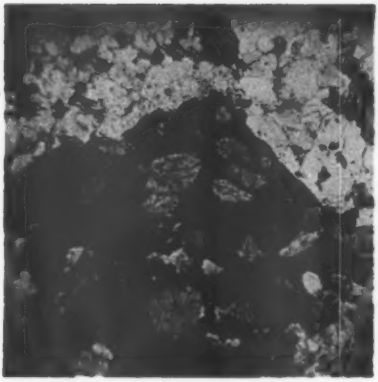

FIGURE 6

SCAPOLITIZATION

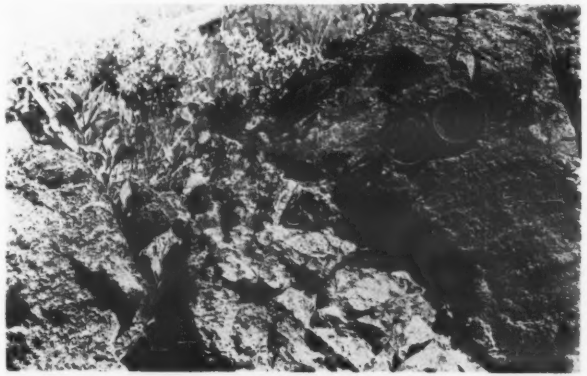

FIGURE 1

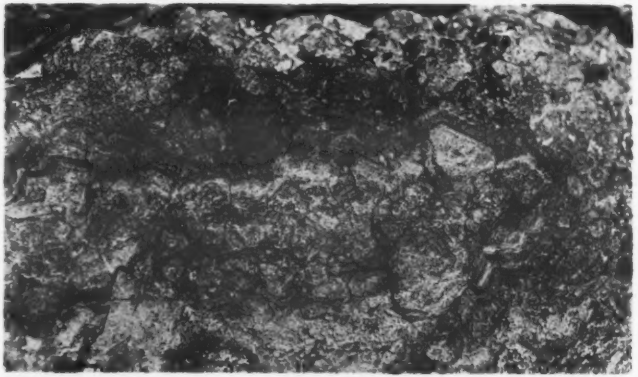

FIGURE 2

FIGURE 3

FIGURE 4

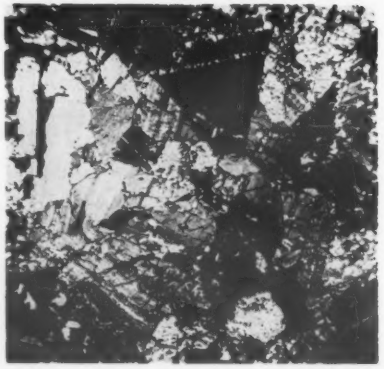

FIGURE 5

ZONED GARNET

FIGURE 2

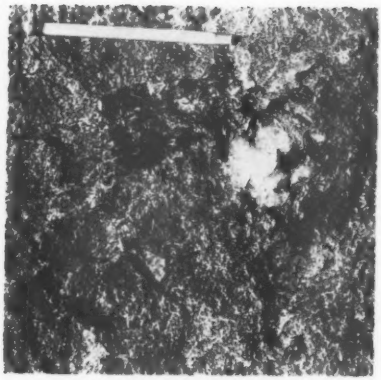

FIGURE 5

FIGURE 4

FIGURE 1

FIGURE 3

LATE MINERALIZATION

tion and appearance. Several feet from the fracture, near more or less pure magnetite-hornblende tactite, the scapolite rock still retains the banding of the magnetite and a large percentage of the hornblende. As the fracture is approached, the rock becomes more massive and a little coarser with less hornblende, until next to the fracture it is a mat of coarse scapolite prisms with included octahedrons of magnetite. Replacement is clearly related to the fracture. The fracture itself may be filled with large prisms of scapolite (Pl. 11, fig. 4).

The relations on Plate 2B, indicate a preference for scapolitization of silicate tactites; they are altered for a greater distance than magnetite or marbles.

Scapolite replaces not only earlier metamorphic types but also aplites. In at least two places (Pl. 2B) bodies of scapolitic rock are somewhat more irregular in shape and penetrate the surrounding rock more than the aplite with which they are associated. For this reason, and on microscopic evidence, the scapolite of these aplites is considered a product of alteration, rather than of primary crystallization, as inferred by Calkins (1909). Calkins' principal evidence for the primary nature of the scapolite was its poikilitic habit (Emmons and Calkins, 1913, p. 125–126); today, however, the formation of such textures by replacement is generally accepted (Shand, 1947a, p. 159).

Refringence measurements on various specimens of scapolite give an average of $N_o = 1.586$, $N_e = 1.544$, which indicates composition near 60 per cent marialite, 40 per cent meionite, or dipyre. Analysis of scapolite rock, replacing tactite on West Algonquin Hill, shows a mol ratio $SO_4:Cl:CO_3::1:7:10$; the excess carbon-

PLATE 12.—ZONED-GARNET TACTITE

FIGURE 1.—CONTACT WITH BANDED TACTITE

Western part of Daly Gulch area. Light-colored rock in lower center is coarse, massive zoned-garnet tactite; it cuts across bedding of darker, grandite-diopside tactite of Red Lion formation. Bedding planes of the latter parallel the Brunton compass.

FIGURE 2.—LARGE ZONED CRYSTALS

Western part of Daly Gulch area; width of picture about 2 feet. Note alternate dark and light zones, and comb structure formed by combination of zones of adjoining crystals. Photograph by J. W. Allingham Jr.

FIGURE 3.—ANISOTROPIC ZONING AND TWINNING

Thin section from south of Douglas Creek, Philipsburg area, ×13, crossed polarizers. Note isotropic cores, and wedges of completely extinguished dodecahedral twins. Zones continuous through all twins of a single crystal, and in many cases between adjacent crystals.

FIGURE 4.—ZONAL ALTERATION

Thin section from same locality; ×26, crossed polarizers. One zone altered to calcite (twinkling gray), and hydrogarnet (black-isotropic).

FIGURE 5.—PERVASIVE ALTERATION

Thin section from same locality; ×13, crossed polarizers. Zoned garnet with diopside and calcite. Note alteration to calcite along single zone near top center.

PLATE 13.—LATE MINERALIZATION

FIGURE 1.—LATE CALCITE VEINS

Back of Trilby tunnel, Southern Cross-Cable area; width of exposure about 3 feet. Vein of calcite and biotite replacing marble. J. W. Allingham, Jr., photograph.

FIGURE 2.—LATE MAGNETITE VEINS

Hand specimen from Pomeroy mine, Southern Cross-Cable area. Vein of late magnetite, with some chlorite (?), zigzags from upper right to lower left. Some magnetite thinly banded with serpentinized olivine.

FIGURE 3.—JOINT PLANE FILLINGS

Vertical joint plane cutting across bedding bands of grandite-diopside tactite in Silver Hill formation, West Algonquin Hill. Filling principally epidote and andradite (dark gray), but near the hand lens calcite, quartz, and magnetite may be seen in center of filling.

FIGURE 4.—EN ECHELON FRACTURE FILLINGS

Small fractures in grandite-diopside tactite, arranged en echelon, and filled with garnet. Width of specimen 3 inches. West Algonquin Hill, Philipsburg area.

FIGURE 5.—VUG FILLINGS

Boston mine, Garnet area. Dodecahedrons of garnet, overgrown with prisms of quartz, in vugs in garnet tactite. Quartz intergrown with lamellar magnetite.

ate is probably a calcite admixture. Evidently a considerable quantity of chlorine was involved in the formation of these rocks.

Microscopic evidence of replacement of most silicates by scapolite is abundant. Garnet, hornblende, and diopside are found in various stages of replacement. Magnetite is less susceptible to replacement by scapolite; as a matter of fact hornblende protected by surrounding magnetite is the last hornblende to be replaced (Pl. 11, fig. 6). In reflected light, scapolite is observed to fill cracks in the magnetite grains, thus there is no doubt that some of the scapolite is later than some of the magnetite. Scapolite rock encloses pods of nearly pure magnetite (Pl. 11, fig. 5), which indicates some movement of iron during deposition of scapolite. But the known earlier origin of a large amount of magnetite in the neighboring hornblende tactite indicates a local redistribution and concentration. Probably very little iron was added, from outside, by the scapolitizing solutions.

The associated calcite is the first instance of carbonate deposition since the early recrystallization.

Origin

The scapolitizing solutions must have been highly alkalic, for the typical scapolite rock contains more soda than does pure pargasite, the principal alkalic mineral of the replaced rocks. The solutions must have contained sufficient carbon dioxide and chlorine to account for the large amount of scapolite and calcite. It must have been capable of replacing the silicate minerals. All these objectives might be accomplished by an alkaline solution of sodium chloride and carbonate, corresponding to the "pneumatolytic scapolitization" of Sundius (1915). The solution must also have been capable of dissolving and reprecipitating the magnetite. In this case the iron may have been transported in aqueous solution, as suggested for the original magnetite deposition. Or, in the presence of the large amount of chlorine, at temperatures above 305°C., it may have moved as Fe_2Cl_6 gas (Stirnemann, 1925). The first-mentioned mode is even more effective at low temperatures, while gaseous transport is more effective at high temperatures. In view of the probable low temperature at this late stage, hydrothermal transport is again more probable.

Two things are unusual in these scapolite occurrences: (1) the close relation to a granitic, instead of a gabbroic, intrusion (Sundius, 1918, p. 98, states, "Vom Granit oder Syenit ist mir kein Fall von Skapolithisierung bekannt."), and (2) the wide variety of minerals and rock types replaced by the scapolite, which in many occurrences replaces only feldspar (Bugge, 1945).

Joint-Plane Mineralization

General Features

Large plane faces covered with coarse crystals of epidote and other minerals are a colorful feature of the West Algonquin Cliffs. These are exposures of joint planes and are found in hard rocks near intrusive contacts, in the Philipsburg, Daly Gulch, Garnet, and other areas. Similar joint fillings are found in the Philipsburg batholith near some of its borders.

Mineralogy

The most common mineral in the joint fillings of nearly all areas is epidote of the pistacite variety. Crystals are elongated up to three-quarters of an inch with sharp domal faces. Dark-red dodecahedrons (truncated by trapezohedrons) of garnet are nearly as common, in the joints in the Silver Hill formation. One measurement on this type of garnet gave a composition of $An_{68}Gr_{18}Al_{14}$. Flattened crystals of magnetite and a filling of calcite, in cases crystallographically continuous over large areas of the joint, are also common. Quartz, amphibole, and scapolite are found in similar joints. All minerals show the one-sided form characteristic of fissure fillings.

In observing a large number of these points, an approximate correlation became apparent between the mineralogy of the joint filling and the type of wall rock. Even though approximate this correlation becomes very striking, when rocks so closely situated as lower Silver Hill hornfels and the tactite just above it show this difference of mineralogy on joint planes that certainly belong to the same set.

Chemical Origin

Wall rock may easily control the mineral deposited by replacement along a vein, from

solutions in that vein. The wall rock may even control the mineral filling vein spaces, by reacting with the solutions. But in this case some of the minerals of the filling are so closely related to the minerals of the wall rock that they must

TABLE 7.—RELATIVE ABUNDANCE OF MINERALS FILLING JOINTS IN VARIOUS TYPES OF ROCK

Wall rock	Joint minerals (in approximate order of abundance)
pure carbonate marble	calcite, tremolite
biotite-calcite marble	not observed
diopside-calcite marble	open
diopside-grossularite-calcite marble	epidote
grandite-diopside tactite	epidote, andradite, magnetite, calcite, quartz
magnetite-hornblende tactite	epidote, magnetite, scapolite
scapolite-hornblende rock	scapolite, epidote
biotite-quartz-orthoclase hornfels; quartzite	actinolite, quartz, magnetite, scapolite
aplite	epidote (very thin coating)
granite, near borders	epidote (sphene at Atlantic Cable Extension)

have had their source in the surrounding rock rather than in a solution of distant origin. The occurrences of epidote and garnet in garnet tactite, and of quartz in quartzite, cannot be coincidence. These joint fillings represent, then, solution and immediate reprecipitation of wall-rock material. Latterly formed joints, in a rock already saturated with solvent material left behind by the earlier phases of metamorphism, would be loci of reduced pressure and reduced solubility. Solutions that permeated the surrounding rock under pressure would have to diffuse (or flow) into the joint openings and deposit some of their dissolved material upon entering the openings.

Local origin of the joint-filling minerals is also seen in smaller, discontinuous fractures in tactites of West Algonquin Hill. These openings are commonly about half to three-quarters of an inch in height, a sixteenth in width, and an inch or more long. They cut across the bedding, but at an angle to the main joint set; they often occur en echelon along a vertical plane. Figure 4 of Plate 13 illustrates their appearance on a hand specimen; note the tapering ends of each fracture. Microscopically the garnets are the same, but that in the vein is definitely crosscutting. Insofar as can be ascertained, these openings do not have communication with other openings or with joint planes. In the grandite-diopside tactite they are filled with garnet, but in the more purely diopside tactite interbedded with it the openings are empty.

Minerals of the main joint fillings were deposited in a rather regular sequence. In some of the tactites of West Algonquin Hill, the sequence is epidote; calcite, garnet, and magnetite, all about contemporaneous; and quartz (Pl. 13, fig. 3). This sequence indicates either: (1) a change, with time, of the material being dissolved by the solutions coming from the wall rock, or (2) the normal sequence of deposition from the solutions in the fractures, with falling temperature. If it is the result of a normal sequence of deposition epidote should not be deposited before garnet; therefore (1) is nearer the truth.

Structural Origin

Of a large number of filled joints examined on West Algonquin Hill, more than half were within 5° of N. 70°E., 60°S. Several of the smaller fractures were about N. 00°E., 60°W. A number of joints carried no mineral deposit, but no consistent attitudes could be found in these, and most of them were rather tight. The axial plane of the Philipsburg anticline lies about N. 30°W., with a steep easterly dip, and the axis dips 25°–30° N. Therefore the main filled joints are almost at right angles to the axis. Despite the fact, already proved, that the whole sequence of intrusion, metamorphism, and mineralization intervened between the folding of the anticline and the filling of the joints, this seems more than coincidence. Possible explanations are: rebreaking along original joints that were still planes of weakness in spite of any healing effects of the metamorphism, or reinstitution of the compressive forces similar to those responsible for the original folding. In view of parallel epidote joint fillings in the main batholith north of Douglas Creek, the latter explanation is tentatively accepted.

Related Occurrences

At the Boston mine, Top O'Deep area, some cavity fillings may be allied in time and mode of origin to the joint fillings just described. Garnet is dark brown, very coarse, shows no signs of bedding, but has many vuggy cavities. Into these cavities grow crystals of garnet in dodecahedrons, quartz showing prism and pyramid faces, and lamellar magnetite (Pl. 13, fig. 5). The quartz and magnetite are both later than the garnet crystals, consistently overgrowing them. The relative age of quartz and magnetite is not so obvious: lamellae of magnetite are commonly intergrown with the crystals of quartz.

Quartz also occurs in cavities in the banded magnetite tactite of the Red Lion, at the Atlantic Cable Extension claim in the Cable-Southern Cross area. These fillings are obviously later than the formation of the surrounding metamorphic rocks, although how much later cannot be estimated.

Veins of calcite, very common in the Cable-Southern Cross area, were not observed in relation to either scapolitization or joint-plane mineralization but are probably related to this phase of the metamorphism. They are best exposed underground, as a white network, cutting the light-colored marble or cutting the darker biotite or magnetite rock which has replaced it in many localities. The carbonate in the veins is coarsely crystalline and is calcite, whether it veins dolomite or calcite.

The late magnetite veins, described above, may belong in this category.

Sulfide and Related Mineralization

Although sulfide mineralization was not investigated in detail, its relationship to silicate and magnetite mineralization was observed. Details concerning later mineralization have been presented by Emmons (1906; Emmons and Calkins, 1913), Pardee (1917), Billingsley and Grimes (1917), and Goddard (1940).

Pyrite occurs in most of the metallic veins of the region, particularly in the gold veins. In the Philipsburg area, the silver-lead-manganese veins, with which it is associated, crosscut all types of metamorphic rock and the intrusive contract and are therefore younger (*e.g.*, the Algonquin vein in Pl. 3). Pyrite is clearly a replacement vein in a specimen from Hope Hill, Philipsburg area (Pl. 8, fig. 4). Pyrite and other sulfides are not found in appreciable quantity in any silicate rocks, and none is found in the post-metamorphic joint fillings, although both of these contain magnetite.

At Top O'Deep, in the Garnet area, pyrite and chalcopyrite occur in small flat-lying quartz veins that cut the contact of tactite and granodiorite. Their relation to the little magnetite of that area was not observed. At the magnetite deposit on Boulder Creek, pyrite occurs as pods and stringers along veins that definitely crosscut magnetite bodies. Microscopic evidence supports the late age of the pyrite.

Emmons (Emmons and Calkins, 1913, p. 229–230) has described all the following minerals, in the Cable mine, as "replacement deposits of contact-metamorphic origin": quartz, pyrite, chalcopyrite, pyrrhotite, arsenopyrite, magnetite, and gold. Although some specimens do contain both pyrite and magnetite, the pyrite usually occurs in separate masses. The large bodies of pyrite-gold ore, such as that in the Lake stope, contain very little magnetite; and the large bodies of magnetite contain only enough pyrite to make the sulfur content objectionable in smelting. The Showers and Cornish stopes, which were very rich in gold, contain a large body of magnetite, but it is restricted to the southwest wall. Throughout the Cable mine, pyrite of the ore bodies is generally associated with coarse anhedral quartz and very coarse calcite, with gold on calcite cleavages. The photomicrograph (Pl. 8, fig. 5) of typical gold ore, from the Showers stope, does not show replacement of pyrrhotite by gold, but elsewhere on the section this relation is evident. (*See also* Emmons and Calkins, 1913, p. 227.) A smoother gray mineral rims the mass of gold. It is silvery white, soft, gives a microchemical test for tellurium, and etch tests indicate tetradymite. In another specimen, from the 6660 level, the sequence: arsenopyrite, pyrite, chalcopyrite is distinctly shown; the latter is space filling.

Just northeast of the Showers stope, a vein

strikes in the usual northwesterly direction, but dips 12°–45° N. This vein is exposed for 400 feet on the 6945 level as a limonite-hematite-clay zone, and in a raise above the northwest drift of the 6865 level as a tabular body of pyrite ½–3 feet thick. This attitude is in marked contrast to the normal vertical aspect of the magnetite bodies.

Most of the evidence indicates that pyrite was deposited later than magnetite, and in a more siliceous environment. This age relation is generally true of ore deposits that contain the two minterals (Schwartz and Ronbeck, 1940, p. 595–597).

No evidence was incompatible with the deposition of sulfides and gold by alkali sulfide hydrothermal solutions or suspensions. Magnetite alters to pyrite in such solutions at 300°C. (Lindner and Gruner, 1939, p. 548). But there are no data concerning the stability of magnetite in such solutions at higher temperatures, so we cannot presently conclude whether the change from oxide to sulfide deposition was due to a lowering of temperature or to a change of sulfide concentration in the depositing solutions.

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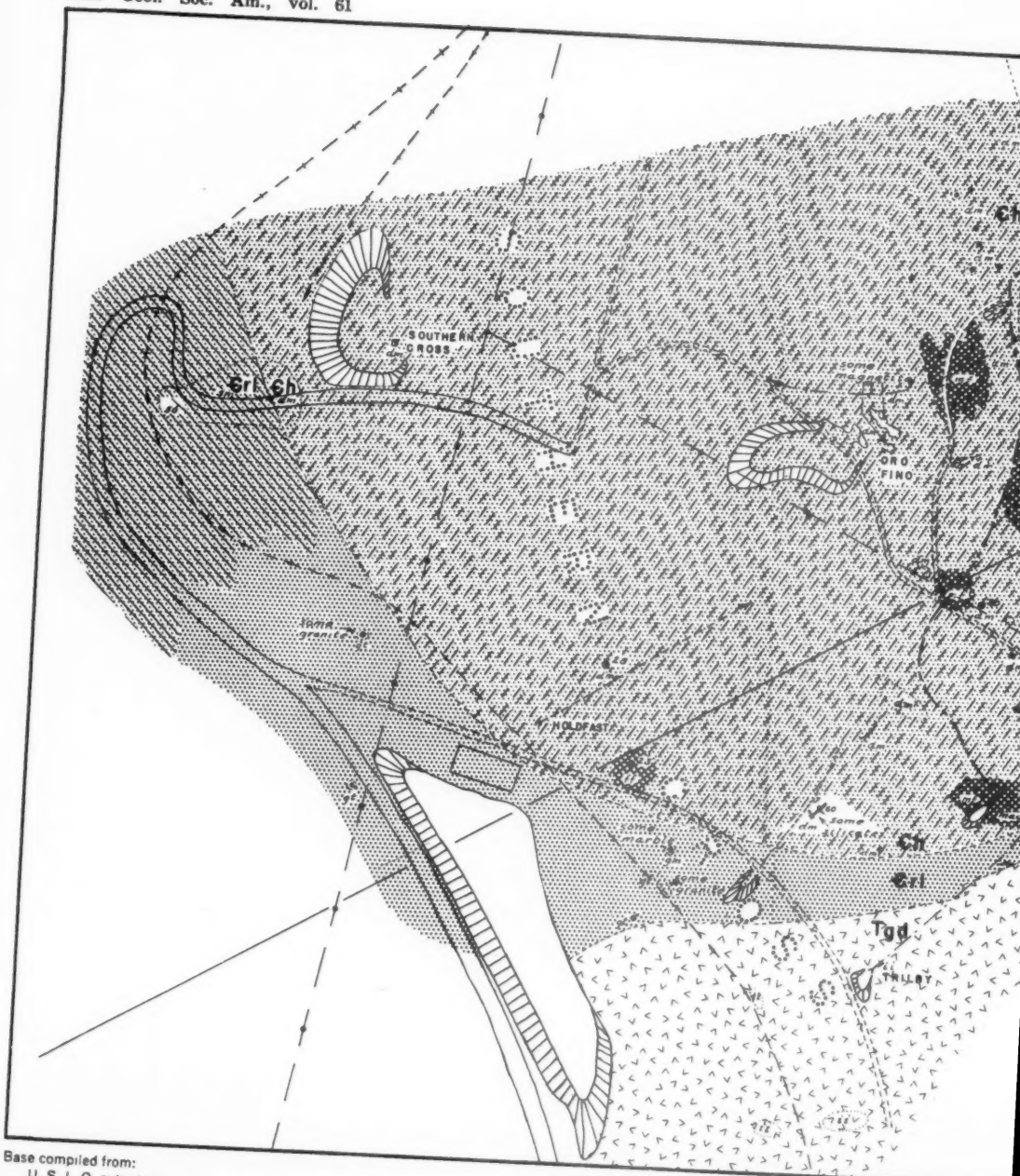

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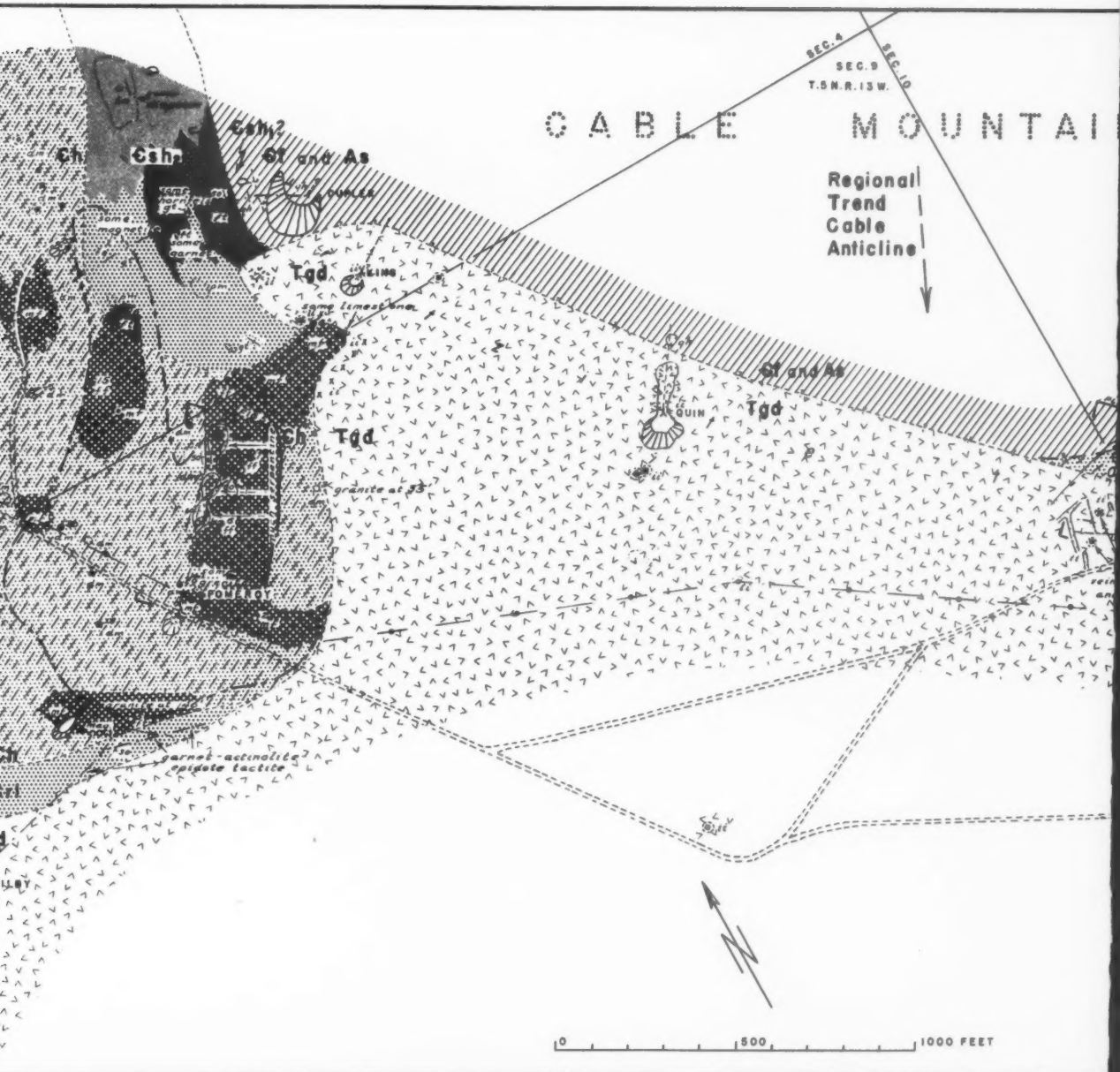
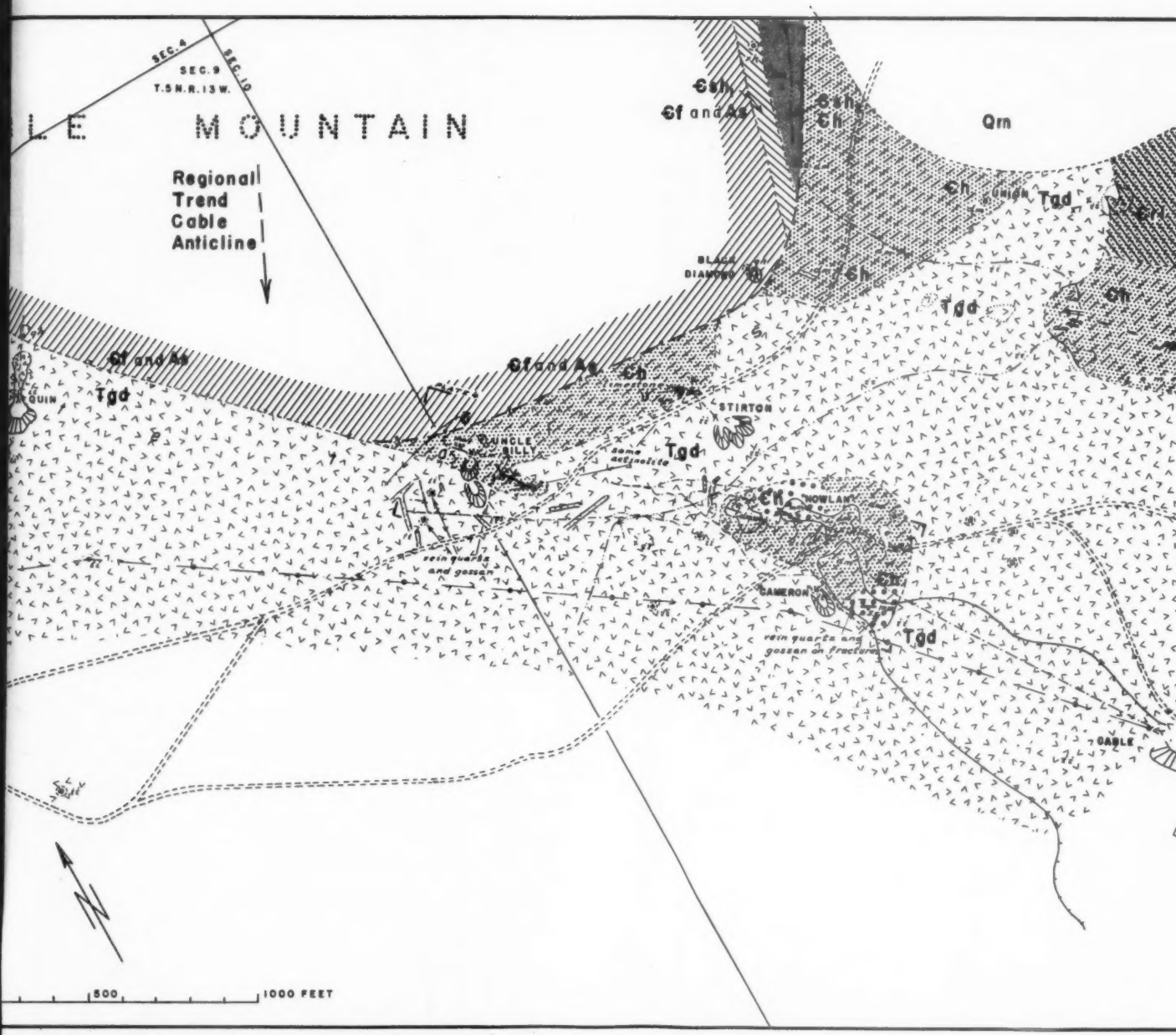


METAMORPHISM IN THE SOUTHERN CROSS - CABLE A



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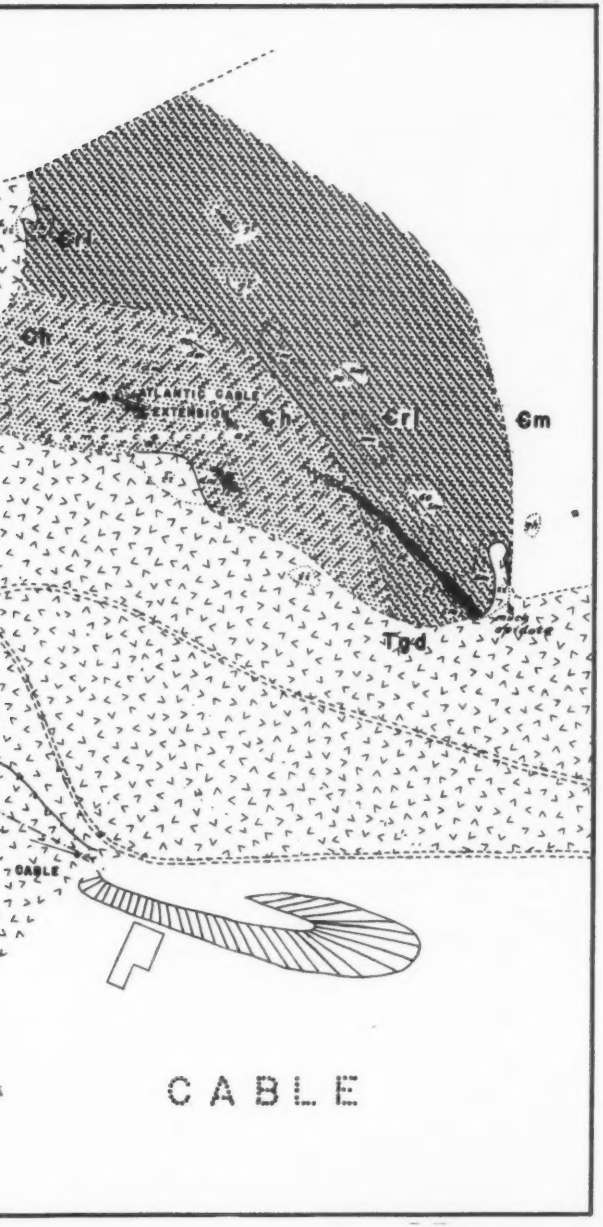

Geology by:
W. T. Holser, 1947-1948
J. W. Allingham, Jr., 1947

EXPLANATION

Lithology

- Dolomite marble
- Coarse calcite marble
- Calcite-silica banded marble
- Biotite-calcite banded marble
- Diopside-calcite banded marble
- Diopside-garnet banded tactite
- Spotted biotite hornfels
- Quartzite
- Magnetite-carbonate tactite
- Magnetite-silicate banded tactite
- Magnetite
- Granodiorite
- Adit
- Magnetic anomaly (USBM 1945)

Stratigraphy

Quat.	Qm	Moraine
Tert.	Tgd	Cable granodiorite
Upper Cambrian	Cm	Maywood formation
	Crl	Red Lion formation
	Ch	Hasmark dolomite
Middle Cambrian	Csh_2	Upper part, Silver Hill formation
	Csh_1	Lower part, Silver Hill formation
	Cf	Flathead quartzite
Belt(ian)	As	Spokane formation

Undifferentiated

- Bedding
- Contact
- Fault
- Dip of vein or ore body
- Shaft
- Caved shaft
- Prospect
- Location recovered Diamond drill holes, with U. S. B. M. number
- Location unrecovered
- Open cut
- Out crop
- Dump
- Surveyed Underground workings
- Caved or inaccessible
- Power line
- Abandoned railroad
- Abandoned ditch

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SUBMARINE GEOLOGY OF BIKINI LAGOON AS INDICATED BY DISPERSION OF WATER-BORNE EXPLOSION WAVES[1]

BY MILTON B. DOBRIN

CONTENTS

	Page
Abstract	1091
Introduction	1092
Acknowledgments	1093
Pekeris normal-mode theory of explosion-wave propagation in shallow water	1093
Experimental procedure	1096
General discussion	1096
Frequency response of recording system	1096
Depth determination	1097
Revision of original refraction interpretation of uppermost zone	1097
Water-wave dispersion observed on Bikini records	1097
General statement	1097
High-frequency dispersion	1097
Parts of lagoon where observed	1097
Method of measuring dispersion characteristics from records	1099
Precision	1100
Data obtained from records	1100
Airy waves	1109
Parts of lagoon where observed	1109
Characteristics of the Airy waves as observed on the records	1109
Use of Airy waves for studying the bottom	1109
Velocity-frequency relationships	1111
Multiple Airy phases	1112
Summary of observed dispersion data	1112
Interpretation of results	1113
General statement	1113
Comparison of Bikini anomalies with those at Barbados and Orinoco Delta	1113
Possible explanations for anomalous dispersion	1113
General statement	1113
Nonuniform water depth	1114
Solid bottom	1114
Vertical changes in velocity	1114
Low-velocity bottom	1114
Horizontal lithology change	1114
References cited	1117

ILLUSTRATIONS

Figure	Page
1. Illustration of group and phase velocities	1094
2. Phase velocity and group velocity (of first mode)	1095
3. Frequency-response curves for channels used to record seismic and water waves at Enyu station	1096
4. Time-distance curves for all profiles	1098
5. Typical plots for obtaining group-velocity, frequency relations	1099
6. High-frequency dispersion cata from shots in area I	1105
7. High-frequency dispersion data from shots in Area IX	1106
8. High-frequency dispersion data from shots in Area VIII	1107
9. High-frequency dispersion observed at Barbados by Worzel and Ewing in 1944	1108
10. Map of Barbados Island, B. W. I.	1116

Plate	Facing page
1. Map of Bikini Lagoon showing locations of shots and detectors	1096
2. Examples of water-wave dispersion on Bikini records: high-frequency patterns	1100
3. Sample records	1101

TABLES

Table	Page
1. Summary of data on high-frequency dispersion	1102
2. Airy phase data	1110

ABSTRACT

The normal-mode theory of Pekeris has been applied to the water-wave dispersion recorded from the under-water explosions set off in the 1946 seismic-refraction shooting in Bikini lagoon. The aim was to ascertain from the dispersion data the velocity of sound in the sediments down to several hundred feet below the lagoon bottom and thereby gain information about the recent sedimentary history of the lagoon. Because of the lack of short shots this information could not be obtained from the refraction results. The principal merit of the

[1] The experimental work upon which the present analysis is based was carried on by the U. S. Naval Ordnance Laboratory and the Oceanographic Section of Joint Task Force One in connection with the 1946 Atomic Bomb tests at Bikini.

dispersion method here is that the long refraction shots supplied some information about the superficial layers.

Information on the first 100 feet or so of sediments below the lagoon bottom was sought from the "high-frequency" dispersion of water waves 50 cps and higher. The data from shots along the edge of the lagoon within 1–2 miles of the reefs and islands appear to be compatible both with normal-mode predictions for a liquid bottom and with refraction indications. From these data it is concluded that the speed within 20 feet of the water bottom averages 1.05 times that in water, with a rapid increase (probably from compaction) to a speed 1.3 times that in water at a depth of about 40 feet. Shots fired somewhat farther from the edge of the lagoon give dispersion curves indicating bottom speeds (on the basis of the Pekeris theory for liquid bottoms) much higher than the refraction data allow. Shots fired in the more central portion of the lagoon give dispersion patterns to which the Pekeris theory does not appear to apply. The author attributes the discrepancy to a lateral variation in sedimentary composition between the edge and center of the lagoon, possibly a downward continuation of the surface change from sand to *Halimeda* debris observed by Emery on going toward the lagoon's center.

The Airy waves (high-amplitude, low-frequency waves at minimum group velocity) give information on the bottom structure to about 400 feet. In virtually all parts of the lagoon their group velocity and frequency are as predicted by the Pekeris theory for a bottom velocity of 6500 ft/sec. This agrees with the first branch of the refraction travel-time curves, which were drawn from very limited data. That the Airy waves fit the theory for shots fired near the center of the lagoon while the higher-frequency waves do not suggests that the lateral change in sediment type within the uppermost layer disappears within a depth of about 100 feet. The change comes at a depth which can only be estimated roughly as 300 feet below sea level and was detected only in the central portion of the atoll.

Introduction

The initial report, by Dobrin, Perkins, and Snavely (1949)[2], on the 1946 seismic-refraction shooting in Bikini Lagoon, carried on in connection with the atomic bomb tests there, discussed the geological information obtained from the seismic waves which traveled along subsurface formations ranging from several hundred to more than 10,000 feet in depth. This information was derived from the time-distance relationships as analyzed by the standard refraction techniques used in geophysical prospecting. Because the shortest shot-detector separation was more than 1000 feet and the next shortest nearly 3000 feet, the refraction data were insufficient to give the exact seismic velocity distribution in the first few hundred feet below the lagoon bottom. Pekeris (1948) has shown that, when explosion waves are generated and received in shallow water, the distribution of seismic velocities can be determined from the water bottom down to several times the water depth by analyzing the frequency dispersion of the waves traveling through the water layer. Pekeris' theory has, therefore, been applied to the water-wave dispersion on the Bikini records in order to obtain information on the bottom composition of the lagoon down to the depth at which the refraction data begin to give reliable indications.

Such information is of particular interest because of the light it may cast on recent deposition of calcareous sediments in the lagoon bottom. Knowledge of this would be valuable in deducing the mechanism by which coral reefs and atolls form. A long-standing problem of geology, this subject has recently attracted renewed attention on account of the scientific investigations associated with the 1946 atomic bomb tests on Bikini atoll.

Direct information on the submarine geology of the lagoons in any of the Pacific atolls is sparse. Skeats (1918) studied the cores from two holes drilled 113 and 144 feet beneath Funafuti lagoon. He found loose uncemented *Halimeda* debris in the first 70 feet and below this 20 feet of cemented calcitic sand consisting of two-thirds of Foraminifera and one-third of *Halimeda*. The last 50 feet of the boring encountered mainly Foraminifera and corals.

As part of the preliminary surveys made in connection with the atomic bomb tests, Emery (1948) took numerous core samples, usually 2 or 3 feet long, from the bottom of Bikini lagoon. Although the samples have not been completely analyzed, several types of bottom material have been found and their distribution determined within the lagoon. In a report on other atolls of the Marshall Islands, Emery,

[2] Henceforth to be referred to as Part I.

Tracey, and Ladd (1949) give similar information about the deeper lagoons, such as Eniwetok.

During the 1946 seismic-refraction survey at Bikini, stringent limitations of time and equipment required concentration on the ground-wave data, and all planning was directed toward this purpose. Water-wave dispersion patterns following the first arrivals were often lost or unreadable because amplitude levels and filter settings most expeditious for bringing out first arrival times did not always show the subsequent dispersion to advantage. Even if instrumental conditions had been ideal, there would have been some question as to the applicability of the Pekeris dispersion theory to the Bikini shooting. The theory calls for uniform water depth and laterally homogeneous elastic characteristics in a bottom which can be treated as a liquid. Neither the bottom topography nor the bottom lithology as far as it is known fulfill these conditions. To a certain extent, therefore, the present investigation tests the extent to which the rather complex bottom conditions in Bikini lagoon conform to the assumptions of the Pekeris theory.

Acknowledgments

The earlier paper on the refraction results of the Bikini seismic survey mentioned the names of many persons who assisted with or participated in the survey. The present author wishes to call attention, in addition, to the contributions made by the co-authors of that paper. Mr. B. Perkins Jr., of the Office of Naval Research, supervised operation of the receiving station on the USS Gillis, from which most of the best dispersion records came. Dr. B. L. Snavely, of the Naval Ordnance Laboratory, designed the entire receiving system and put it into working order under most unfavorable field conditions in time for use in the seismic operations.

In the present analysis of the dispersion data, the author wishes to acknowledge the contribution made by Prof. Maurice Ewing in suggesting the dispersion study and in giving valuable consultation throughout the course of the work. He also wishes to thank Dr. Frank Press for reading the manuscript and for many helpful discussions.

Pekeris Normal-Mode Theory of Explosion-Wave Propagation in Shallow Water

During the course of some war-time research on the propagation of explosion sounds in shallow water, Worzel and Ewing (1948) observed that water waves from an under-water explosion exhibit a variation of velocity with frequency in which each wave length has a characteristic speed dependent on the water depth but independent of the distance traveled. In experiments conducted in a number of widely separated areas along the U. S. east coast, the West Indies, and the lesser Antilles, they noted that in each locality the velocity vs. frequency relation was the same for all shots within any given area of approximately constant water depth.

Upon learning of these results, Pekeris (1948), who had for some years been giving attention to the propagation of compressional waves in layered liquid media, worked out a theoretical explanation for the experimental data. His theory, based on the use of normal modes, not only explains most of Worzel and Ewing's observations but also predicts additional features which he has subsequently identified on their records. Press and Ewing (1948a; 1948b) have extended Pekeris' treatment, which he developed originally for layered liquids where the deeper layers have higher speeds, to cover the case where the liquid layer is underlain by a solid medium, as well as that where the uppermost layer lies over another liquid layer having a lower speed. In more recent papers (*see* Ewing and Press, 1950, and Press *et al.*, 1950), the same authors have applied dispersion theory to study layering under the ocean basins.

Although the complete derivation of the normal-mode theory presented in these papers will not be reproduced here, a brief qualitative summary may facilitate understanding of the dispersion results from Bikini. Basic to any discussion of dispersion is the concept of group and phase velocity. When an explosion is detonated under water, the initial pressure pulse traveling away from the source can be resolved by Fourier analysis into sinusoidal components having all frequencies from zero to infinity. For wave lengths no shorter than about one tenth

the water depth, dispersion will be observed, and each frequency will travel through the water at a different speed. The initial pulse will now spread out into a train of almost sinusoidal waves having frequencies continually varying with their position in the train.

In such a train, any given wave cycle will change in length as the wave moves away from its source. If an observer follows the crest of such a wave he finds it has, at any point, an instantaneous velocity, c, which varies with the frequency. This is defined as the *phase velocity* (Fig. 1). If, at each distance, he selects a pair of crests separated by a given interval (*i.e.*, constant wave length) he finds that this wave length is propagated at a characteristic speed depending on frequency. This speed, known as the *group velocity*, U, is obtained directly from the record by dividing the shot-detector distance by the total travel time of the wave in question.

The phase velocity of a compressional explosion wave transmitted without damping through shallow water will vary inversely as the sine of the angle made with the vertical by the multiply reflected wave paths allowing maximum phase reinforcement for each frequency. This angle will be different for each wave length and for each allowed mode of transmission at that wave length. Thus for any mode (usually only the first need be considered if the proper measurement techniques are employed) the phase velocity and group velocity will depend on the wave length, and the water waves will exhibit dispersion. For a given water depth, the precise form of the dispersion curve will depend only on the density of the bottom material and its elastic characteristics. It is thus possible to study the bottom structure from the dispersion of the compressional waves propagated through the overlying water layer.

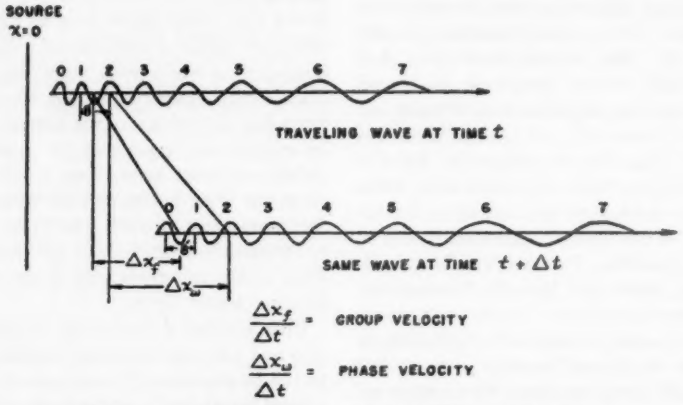

SOURCE
x=0
0 1 2 3 4 5 6 7
TRAVELING WAVE AT TIME t
0 1 2 3 4 5 6 7
SAME WAVE AT TIME t + Δt
Δx_f
Δx_w
Δx_f / Δt = GROUP VELOCITY
Δx_w / Δt = PHASE VELOCITY

FIGURE 1.—ILLUSTRATION OF GROUP AND PHASE VELOCITIES

The two wave patterns represent successive cross sections of a traveling wave, such as a surface water wave; the second is a short time, Δt, after the first. Horizontal co-ordinate is distance from source.

Figure 2 shows a number of curves of c and U vs. frequency, f for a water layer with a speed of c_1 and a thickness H underlain by a liquid bottom of speed c_2. These were calculated on the basis of Pekeris' normal mode theory. The frequencies are expressed in the dimensionless form $\frac{fH}{c_1}$, and the velocities in the form c/c_1 and U/c_1. Each curve represents a different c_2/c_1 ratio, covering the range from 1.05 to 3. If the upper medium is water with a sound velocity of 5000 ft/sec and the lower medium is sedimentary rock (considered as a liquid), these curves, extending from c_2 = 5250 ft/sec to c_2 = 15000 ft/sec, should be valid for virtually all homogeneous bottom sediments, assumed to be liquid, for which the sound velocity is greater than that in water.

A most significant feature of the group-

velocity curves of Figure 2 is the minimum observed in each. The higher the c_2/c_1 ratio the lower is the frequency at which this minimum occurs. Havelock (1924) shows that any maximum or minimum in the group-velocity vs. frequency curve will result in a situation in which components having adjacent frequencies will travel at the same velocity and because of their reinforcement will give an amplitude maximum. The amplitude peak at the frequency of minimum-group velocity is readily identified on most oscillograph records of shallow-water explosions and is called by Pekeris the *Airy wave*.

The first "water wave" to arrive after the initial ground wave should have, according to Figure 2, a frequency approaching infinity; subsequent water waves should be of successively lower frequency. The water-wave arrivals, represented by the portion of the group-velocity curve to the right of the minimum, should be superimposed on the lower-frequency ground waves, represented by the portion of the curve left of the minimum and having a frequency which increases with time. At the minimum-group velocity the water wave and ground waves coincide and give rise to the Airy peak.

Insofar as one can consider the unconsoli-

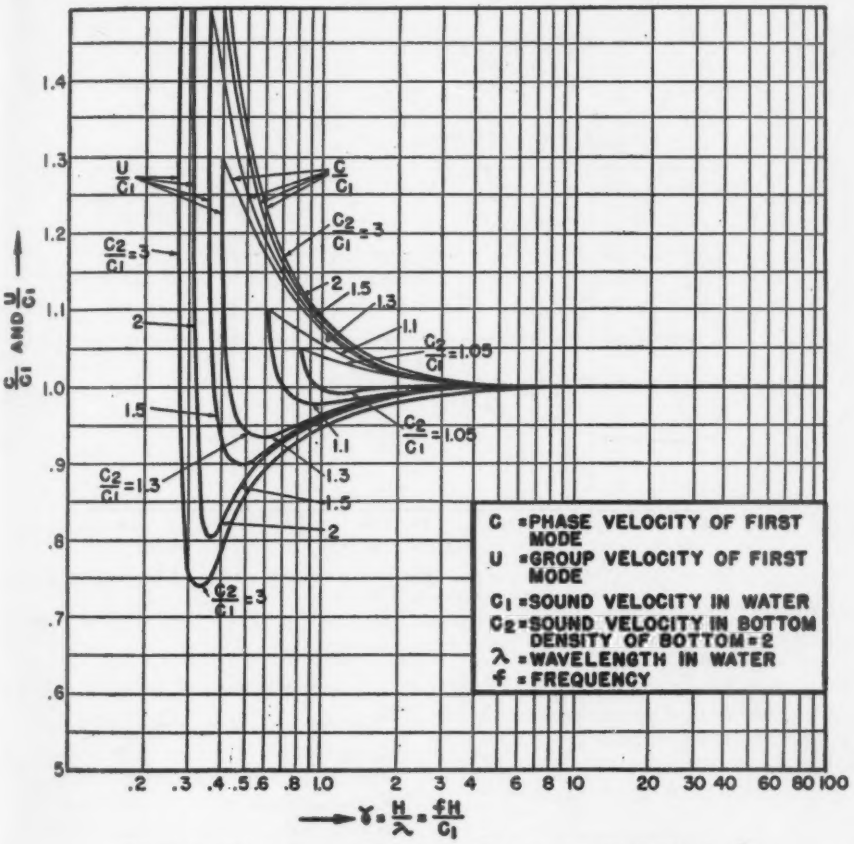

C = PHASE VELOCITY OF FIRST MODE
U = GROUP VELOCITY OF FIRST MODE
C1 = SOUND VELOCITY IN WATER
C2 = SOUND VELOCITY IN BOTTOM
DENSITY OF BOTTOM = 2
λ = WAVELENGTH IN WATER
f = FREQUENCY
U/C1 AND C/C1
γ = H/λ = fH/C1

FIGURE 2.—PHASE VELOCITY AND GROUP VELOCITY (OF FIRST MODE)
In water layer of thickness H underlain by liquid bottom with sound velocity c_2/c_1 times as great (after Pekeris, 1948).

dated bottom material in Bikini lagoon to have the elastic properties of a liquid as well as a uniform sound velocity greater than 5000 ft/sec, the speed of sound in it and, hence, its elastic properties should be deducible by comparing the observed group velocities with the theoretical curves of Figure 2.

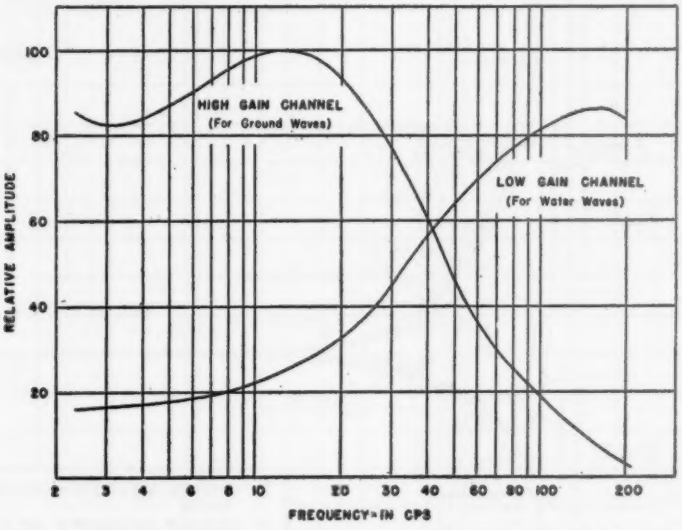


FIGURE 3.—FREQUENCY-RESPONSE CURVES FOR CHANNELS USED TO RECORD SEISMIC AND WATER WAVES AT ENYU STATION

Airy phases were recorded on high-gain channel. High-frequency water waves recorded on low-gain channel.

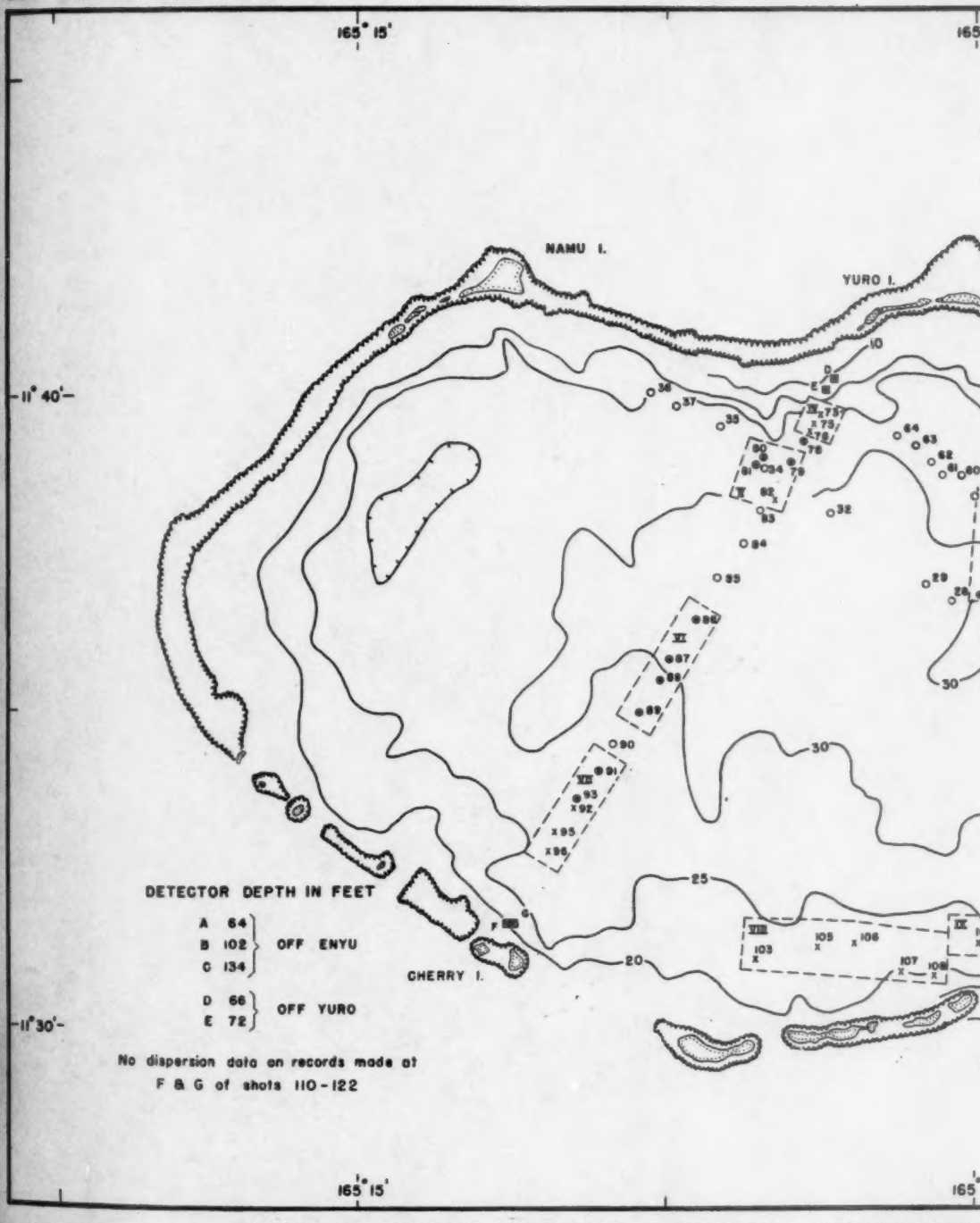
EXPERIMENTAL PROCEDURE

General Discussion

The technique of shooting and recording adopted for the seismic study of Bikini Atoll has been described in Part I. The Mk1 hydrophone detectors, as pointed out there, were usually on the sandy terrace near shore, while the shots were mostly farther from shore in the deeper portion of the lagoon. Both were always at the water bottom. Plate 1 shows locations of the shots giving dispersion data as well as those of all detectors. There is about 70 feet of variation in water depths at the three detectors off Enyu Island. Cables connected these detectors to recording instruments on the island, but the outputs of the other detectors were recorded by equipment of a similar type on the USS GILLIS. All charges giving usable dispersion data were 300 lbs. (Mk6 naval depth charges).

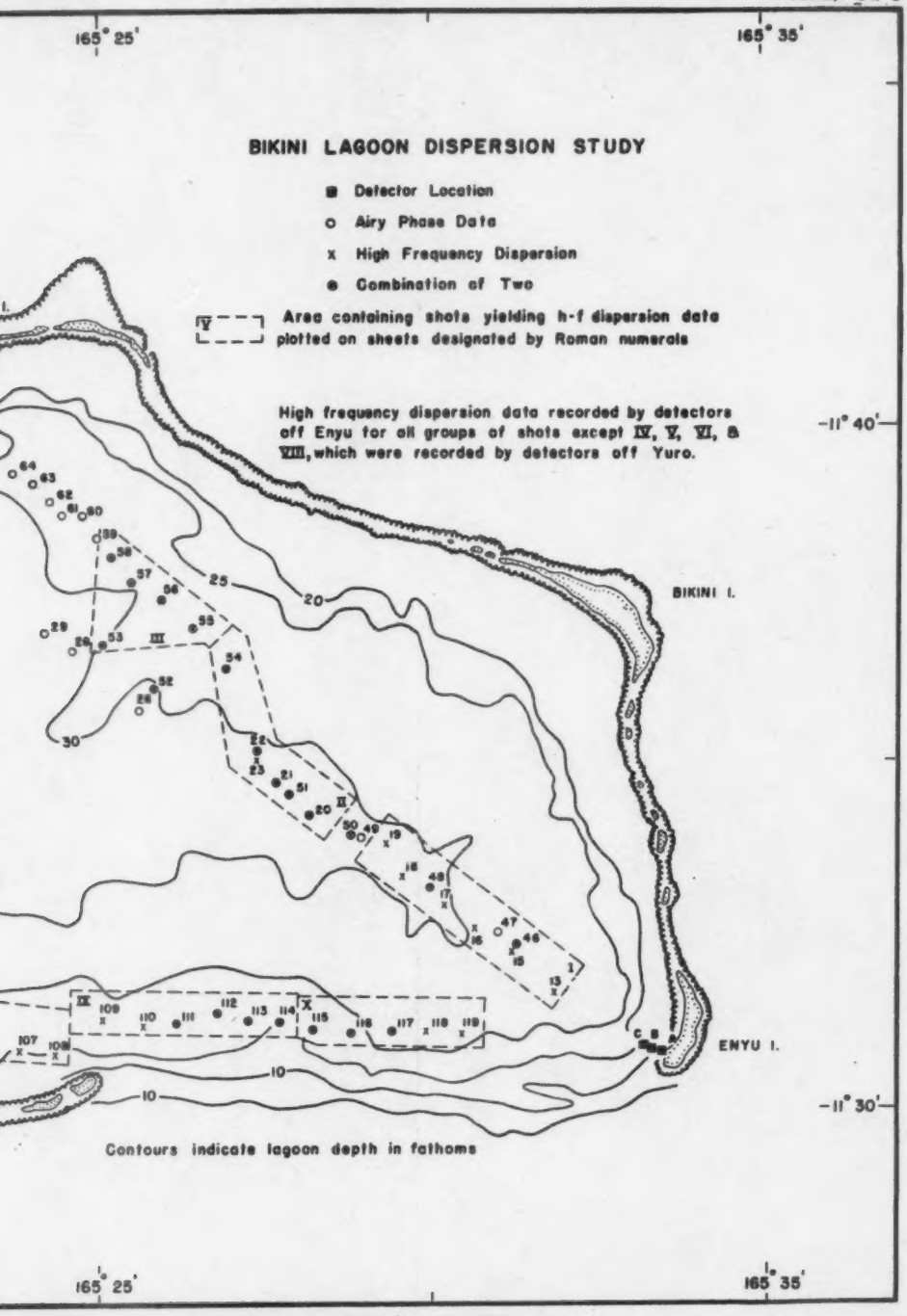
Frequency Response of Recording System

The Mark 1 system employed at both recording stations responds, when no filters are introduced, to frequencies from zero to about 300 cps, although there is a substantial fall-off beyond 200 cps. The refraction survey required accurate recording of ground waves with a frequency in the 10 to 20 cps range and also of the earliest arriving water waves. The channel for detecting the long-period ground waves had a low-pass filter and a high-gain setting. The channel for recording the water-wave first arrivals contained a high-pass filter. Figure 3 gives typical overall response curves for the two traces as recorded by the system on Enyu Island. The system employed on the USS GILLIS is similar in design, although no frequency-response curve could be made for it. The low-pass filter cuts off at such a low frequency that it tends to eliminate the high-frequency arrivals associated with the first



MAP OF BIKINI LAGOON SHOWING LOCATIONS

Shots giving dispersion data divided into zones indic


165° 25'
165° 35'
BIKINI LAGOON DISPERSION STUDY
Detector Location
Airy Phase Data
High Frequency Dispersion
Combination of Two
Area containing shots yielding h-f dispersion data plotted on sheets designated by Roman numerals
High frequency dispersion data recorded by detectors off Enyu for all groups of shots except IV, V, VI, & VIII, which were recorded by detectors off Yuro.
-11° 40'
BIKINI I.
ENYU I.
-11° 30'
Contours indicate lagoon depth in fathoms
165° 25'
165° 35'

ATIONS OF SHOTS AND DETECTORS

zones indicated by Roman numerals

mode. The "water-wave" channel did not generally give a satisfactory trace for dispersion studies because of its tendency to admit higher modes which complicate the dispersion pattern. Airy phases were transmitted quite adequately by the low-pass "ground-wave" channel since their frequencies were all near 20 cps.

Depth Determination

Since none of the shooting ships were equipped with a fathometer and since available lead lines were considered unreliable in 30 fathoms when dropped from a ship underway, water depths were ascertained from chart H. O. 6032. The soundings printed there were made by Japanese hydrographers. A check with a recent preliminary and presently restricted chart made by Emery on the basis of his more detailed fathometer survey of 1946 shows that depths interpolated from the chart are sufficiently accurate for the present study.

Revision of Original Refraction Interpretation of Uppermost Zone

In Part I, the refraction time-distance curves were resolved into three segments having average inverse slope of 7000, 11,000, and 17,000 ft/sec respectively. On the basis of indications from the present study the time-distance data have been re-examined, and it now appears that the 700 ft/sec zone can be equally well represented by two layers. Figure 4 shows the first portions of the time-distance curves for the four profiles of the refraction survey. Two segments now replace the original 7000 ft/sec segment. Data points are not available close enough to the shot locations to specify the inverse slope of the first segment precisely, but they do establish a *maximum* speed for it of 6500 ft/sec.[3] The second segment has a slope corresponding to a speed of 7500 ft/second.

Using the standard formula for the depth to the interface in the two-layer case (Part I, p. 812) and substituting .06 sec. for T_{i1}, 6500 ft/sec for V_1, and 7500 ft/sec for V_2, one gets a thickness of 390 feet for the "6500 ft/sec" zone. Using the second formula on the page for the thickness of the 7500 ft/sec layer underlain by an 11,000 ft/sec zone and introducing an average T_{i2} of 0.45 sec., a thickness of 1800 feet is obtained for the second layer. The sum of the two thicknesses, about 2200 feet, is very little different from the thickness originally obtained for the single 7000 ft/sec layer.

The 6500 ft/sec speed is the maximum that can be expected to a depth of 400 feet below the lagoon bottom if the material above this depth is considered homogeneous. On the basis of compaction theory, however, one would expect the formations near the surface to have a lower velocity than this so that the sound velocity would be somewhat higher than 6500 feet just above the 7500 ft/sec discontinuity. It is clear, however, that any dispersion data that indicate a sound velocity within 400 feet of the water bottom substantially higher than 1.3 to 1.5 times that in water are in disagreement with the refraction data.

Water-wave Dispersion Observed on Bikini Records

General Statement

About half the shots exploded along the four refraction profiles yielded records with measurable frequency dispersion in the interval directly following the first water-wave arrival. As observed by Worzel and Ewing and as explained by Pekeris' theory, the group velocities for frequencies higher than the Airy frequency increased with increasing frequency, the highest-frequency water wave thus arriving first. The last water waves to arrive were the Airy waves, having the lowest frequency and traveling at the minimum-group velocity.

High-Frequency Dispersion

Parts of lagoon where observed.—High-frequency dispersive water waves were recorded 3 to 14 miles from the shots by the Enyu detectors for shots along the Enyu-Namu, Enyu-Yuro, and Enyu-Cherry profiles and by detectors recording on the USS Gillis off Yuro for shots along the Yuro-Cherry profile. Of all shots within this range only those recorded from the Gillis off Cherry and shot along the

[3] This value has significance only to the extent that a *single* homogeneous layer can be assumed to overlie the 7500 ft/sec zone.

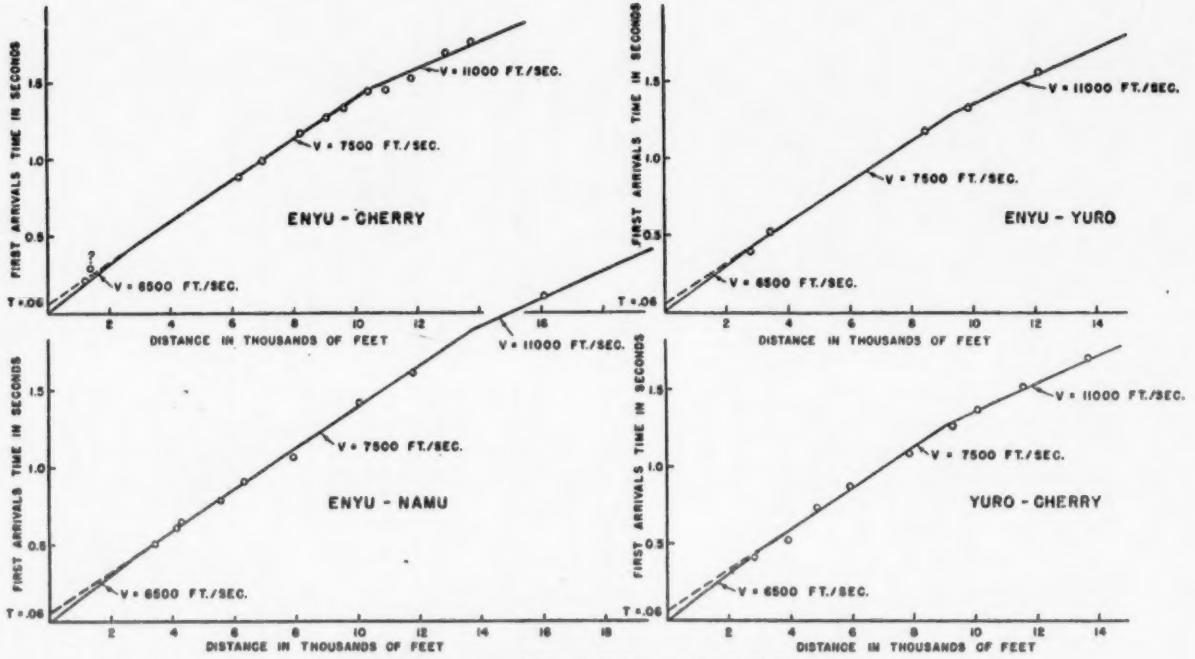
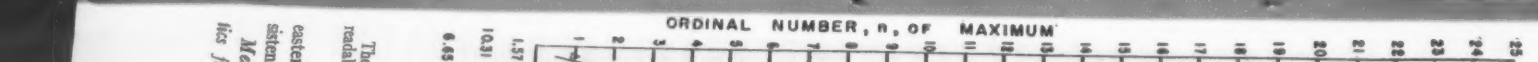


FIGURE 4.—TIME-DISTANT CURVES FOR ALL PROFILES

Showing partition of original "7000 ft/sec zone" into two separate layers. Maximum possible speed in uppermost layer is 6500 ft/sec, but true value may be lower.



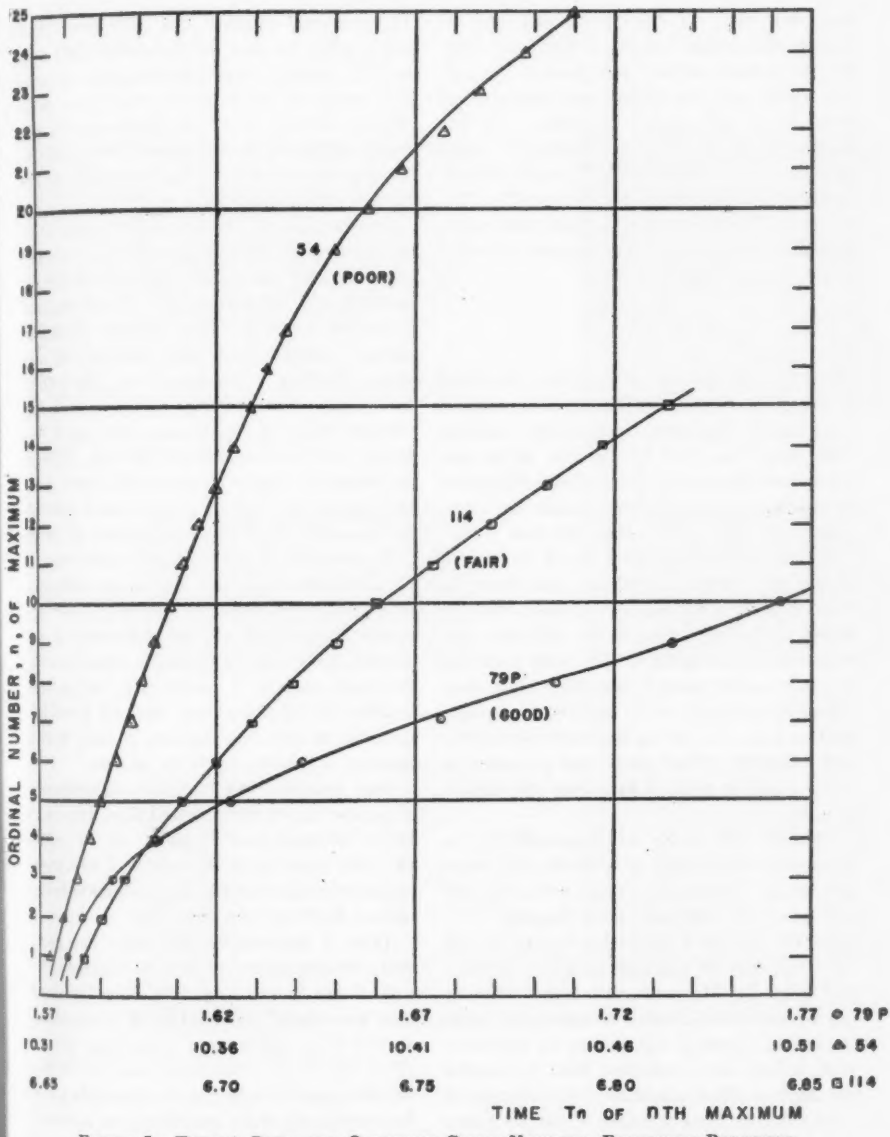


FIGURE 5.—TYPICAL PLOTS FOR OBTAINING GROUP VELOCITY, FREQUENCY RELATIONS
The quantity T_n is time of peak in water wave, n the ordinal number of the peak counting from the first readable water-wave cycle.

eastern part of the Enyu-Cherry profile consistently failed to show dispersion.

Method of measuring dispersion characteristics from records.—The technique of obtaining the group velocity vs. frequency relations from the records was similar to that employed by Pekeris (1948, p. 8). Times of successive peaks representing successive water-wave cy-

cles were read on the record and plotted against the ordinal number of the peak (Fig. 5). A smooth curve was drawn through the points, and the period was measured at intervals of 0.01 second by determining the slope of the curve. The group velocities corresponding to each arrival time, T, on the records were expressed in terms of the fraction $(T - T_o)/T_o$, where T_o is the time of the first water-wave arrival. The frequencies were expressed in terms of the dimensionless quantity

$$\gamma = \frac{H}{\lambda} = \frac{H}{c_1 \Delta t}$$

where H is the average water depth, c_1 the speed of sound in water, and Δt the crest-to-crest wave period. The speed c_1 was always taken as 5000 ft/sec.(*See* Part I.) Because of the unsatisfactory filtering the most serious difficulties in reading the records were caused by interference of modes higher than the first.

The original records from which the curves of Figure 5 were taken are reproduced in Plate 2. Both good records and poor ones are shown to illustrate some of the difficulties encountered in the analysis. The wide variation in record quality made it desirable to introduce a grading system in which records were evaluated as *A*, *B*, or *C* on the basis of their validity and reliability. These grades are presented in Table 1 and in plots of data from the various records.

Precision.—In using the group-velocity vs. frequency relationships to determine the structure of the bottom the requirements for precision increase with increasing frequency, because the dispersion curves for various bottom velocities tend to converge at higher frequencies (Fig. 2). Moreover, the higher the frequency, the more difficult it becomes to measure it precisely because the change in frequency with arrival time is greatest when frequencies are highest. The graphical determination of period is thus least accurate in the frequency region where the least error can be tolerated. The determination of group velocities is also least precise at high frequency because T and T_o are closest together here. The quantity T can usually be read to the nearest .001 sec, but T_o, without a very high-frequency or rectified trace, will generally be read from about .002 to .010 sec. late. The significance of this error decreases as the interval $T - T_o$ increases, and becomes negligible usually at frequencies corresponding to $\gamma < 5$.

Another source of uncertainty particularly troublesome in analyzing the Bikini dispersion data involves the proper value to use for the water depth. The normal-mode theory assumes a uniform depth of water over the whole trajectory between shot and detector. In the Bikini shooting, as the map shows, the bottom topography was far from uniform, and the average value of depth along the wave trajectory had to be used in the analysis. There is no reason to expect a systematic error from this source that would not be much reduced by averaging data from a number of shots.

To summarize, the only systematic error is that in reading T_o, which will always be slightly late. This error will give rise to a lower apparent group velocity; the difference is observable only when γ is greater than about 5. A certain amount of scatter is to be expected in plots of dispersion data, but, for lower frequencies at least, the average of data from a number of records should be reliable.

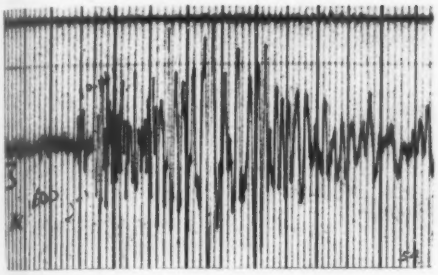
Data obtained from records.—High-requency dispersion curves were plotted from data on 65 traces recorded from 55 shots. On 10 records the data were taken from each of two traces representing the output of two separate hydrophones receiving the same shot.

Table 1 summarizes the data. The group velocities are presented both as functions of γ and of the frequency f. The "average depth over wave path" was used for H in computing γ. The U/c_1 values were taken from plots of $(T - T_o)/T_o$ vs. γ and were based on the best smooth curve through the observed data points. In general only slight smoothing was necessary.

To evaluate the significance of the dispersion data in Table 1, it is necessary to compare the group velocity-frequency relations with those

PLATE 2.—EXAMPLES OF WATER WAVE DISPERSION ON BIKINI RECORDS: HIGH-FREQUENCY PATTERNS

Note increase of period with time after first arrival, indicating decrease of group velocity with increasing period.

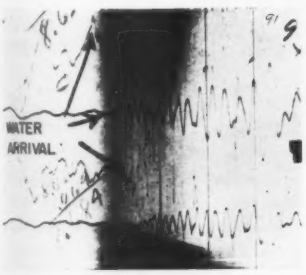


SHOT 54 DET. B GRADE C

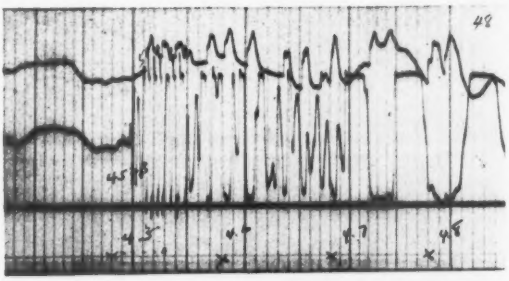
SHOT 88 DETS. D&E GRADE A

SHOT 87 DET. D GRADE A

SHOT 79 DET. D GRADE A

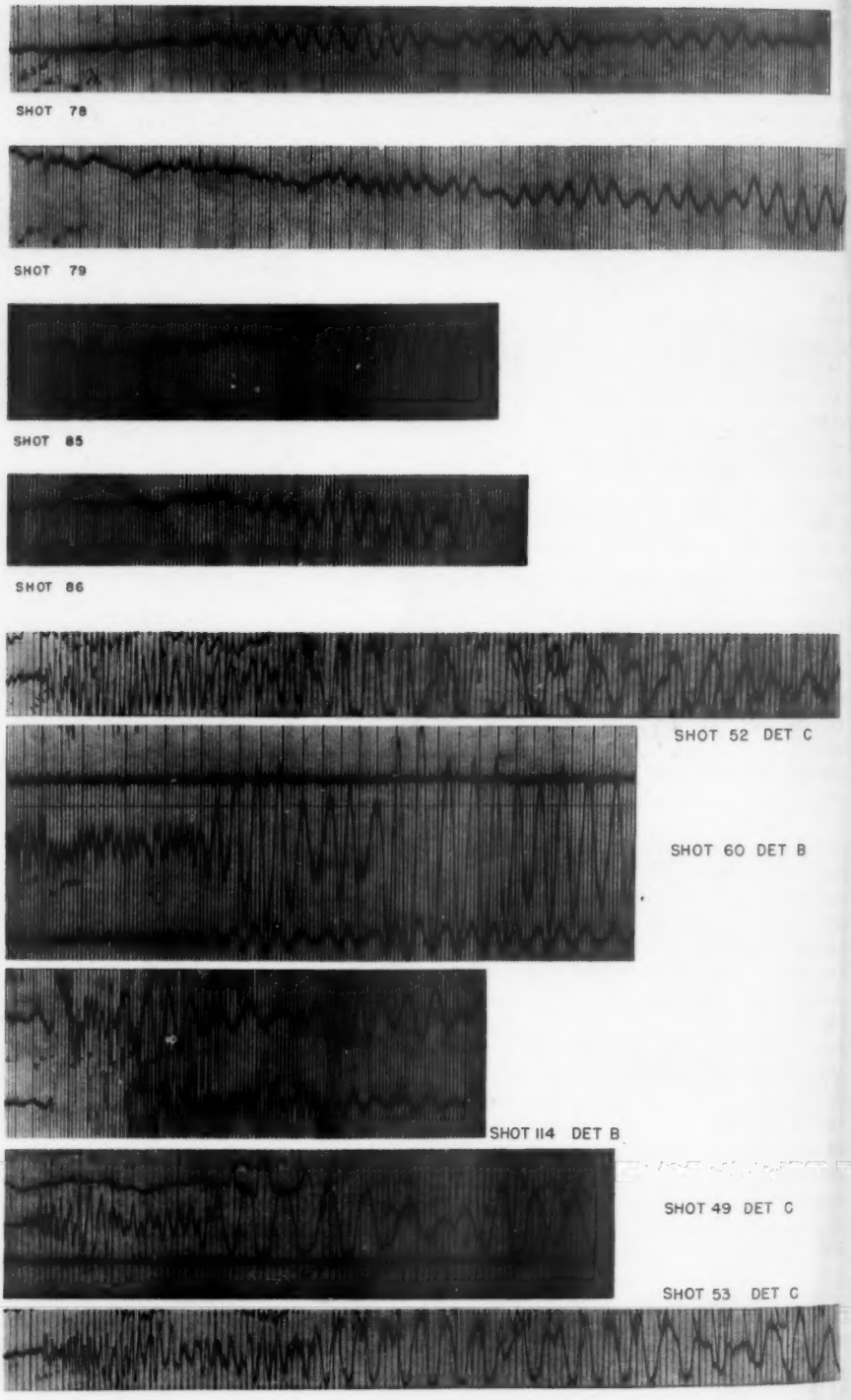


SHOT 91 DETS. D&E GRADE A



SHOT 48 DET. C GRADE C

EXAMPLES OF WATER-WAVE DISPERSION ON BIKINI RECORDS: HIGH-FREQUENCY PATTERNS



SAMPLE RECORDS

predicted by the Pekeris theory. This can be done by plotting the data on the same coordinates as a set of theoretical curves such as those in Figure 2. To increase the resolution of the high-frequency data, however, the group velocities are indicated in terms of the dimensionless fraction $(T - T_o)/T_o$, and the reference curves of Figure 2 have been replotted on this basis. This representation is the same as that used by Pekeris for plotting the dispersion data from Worzel and Ewing's records.

Figures 6 to 8 show the data thus plotted for some representative sets of records indicating high-frequency dispersion. The shots were grouped for purposes of comparison into zones of uniform dispersion characteristics. The boundaries of these zones along the various profiles are indicated on the map (Plate 1), and the zones themselves are specified by Roman numerals. A separate dispersion plot (of the type shown in Figs. 6–9) was made for each zone.

For each plot, all poinst lay, within the limits of uncertainty discussed above, inside narrow zones parallel to the theoretical "group-velocity" curves. There appears to be no relation between the dispersion characteristics and the shot-detector distance. The points on each plot could usually be represented by a single curve constituting the "axis" of the elongated cluster. Within each area represented by a separate dispersion plot, the group velocity depends only on the frequency and depth of water. This conforms to the conclusion originally reached by Worzel and Ewing (1948) on the basis of their 1943–1944 observations.

There are conspicuous differences in the dispersion patterns exemplified by each plot. If the normal-mode theory is applicable the velocity distribution of the bottom material should be ascertainable by noting where the points lie with respect to the curves for various c_2/c_1 ratios. The depth below the water bottom to which the data apply depends on frequency, and water depth is given by the curves labeled "Depth of Penetration"[4] plotted above the dispersion curves. The lower the group velocity (the higher the $(T - T_o)/T_o$ ratio) for a given frequency, the greater will be the corresponding c_2/c_1 value. The values of c_2/c_1 represented by the theoretical curves cover the range from 1.05 to 3, and, since refraction data suggest that the speed in the first 400 feet averages no more than 1.3 times the speed of sound in water, one would expect the average of the observed data points to lie in the vicinity of the line for $c_2/c_1 = 1.3$.

It was therefore surprising to find that for many of the areas, such as that represented in Figure 6, the data points lay well outside the range of the theoretical curves, the group velocities being so low that no possible c_2/c_1 value could apply. These areas were mostly in the deepest portion of the lagoon. Since the known error in T_o should have given a discrepancy in the direction of a higher rather than a lower group velocity and since no other experimental error could be discovered which would affect the group velocities systematically, it was concluded that the conditions in the anomalous areas do not comply with the assumptions of the theory.

Group velocities lower than the theory permits were observed at two of the locations where Worzel and Ewing did their shooting, namely Barbados and Orinoco (deep). Figure 9 is a plot of the dispersion data from six of their records made at Barbados. The departure from the permissible range is in most cases greater than for the anomalous Bikini records.

In several of the Bikini areas (mostly nearer the lagoon edge) for which separate dispersion plots were made the points lie within the theoretical range, but the apparent c_2/c_1 values are consistently higher than the maximum suggested by the refraction data. Most of the points from

[4] Pekeris defines the "Depth of Penetration" as the depth above which 99 per cent of the explosion wave energy is transmitted.

PLATE 3.—SAMPLE RECORDS

FIGURE 1.—AIRY WAVES ONLY

High-frequency waves apparently absorbed along the travel path or cut out by filter system in recording equipment.

FIGURE 2.—HIGH-FREQUENCY AND AIRY WAVES ON SAME RECORD

Note multiple Airy peaks.

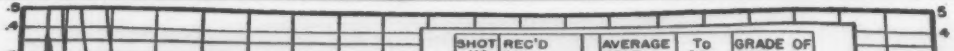
TABLE 1.—SUMMARY OF DATA ON HIGH-FREQUENCY DISPERSION

Shot no.	Rec'd by Detector	Rec. Grade	Shot Depth ft.	Av. Depth water ft.	$\theta°$	$\gamma = 1.5$		$\gamma = 2.0$		$\gamma = 3.0$		$\gamma = 4.0$		$\gamma = 5.0$		$\gamma = 6.0$		$\gamma = 7.0$	
						freq cps	$\frac{V}{c_1}$	freq cps	$\frac{V}{c_1}$	freq cps	$\frac{V}{c_1}$	freq cps	$\frac{V}{c_1}$	freq cps	$\frac{V}{c_1}$	freq cps	$\frac{V}{c_1}$	freq cps	$\frac{V}{c_1}$
13	A	C	150	150	2.376	50	.963	67	.976	100	.9856	133	.9904	167	.9922	200	.9936		
15	A	B	170	159	3.223	47	.954	63	.972	94	.988	126	.9928	157	.9947	188	.9958	220	.9965
16	A	B	166	162	4.011			62	.974	93	.9845	123	.9907	154	.9942	185	.9962	216	.9972
17	A	C	172	162	4.922					93	.986	123	.9910	154	.9935	185	.9952	216	.9965
18	A	C	172	164	5.644			61	.9781	91	.987	122	.9920	152	.9941	183	.9954	213	.9965
19	A	C	170	168	6.133					89	.989	119	.9928	149	.9950	178	.9963	208	.997
20	A	C-	170	172	7.701			58	.9825	87	.9865	116	.9901	146	.9929	174	.9947		
21	A	C-	168	160	8.607	47	.976	62	.9810	94	.9865	125	.9895	157	.9923	188	.9940	219	.9951
22	A	C-	126	160	9.071			62	.9780	94	.9875	125	.992	157	.9943	188	.9959	219	.997
23	A	C-	1151	170	9.337	44	.976	59	.9800	88	.987	118	.991	147	.9934	176	.9952	206	.9966
46	A	B	166	171	3.297	44	.976	59	.983	88	.989	117	.9928	147	.9955	175	.9968		
"	C	B	"	"	2.830	44	.976	59	.981	88	.987	117	.9920	147	.9949	175	.9962		
48	A	C	175	168	4.938			60	.984	89	.9902	119	.9927	149	.9945	179	.9968		
"	C	C	"	"	4.523					89	.979	119	.9865	149	.9907	179	.9932		
50	A	B-	162	174	6.819	43	.969	58	.979	86	.9905								
"	C	B-	"	"	6.353			58	.9845	86	.9922	115	.9946	144	.9958	172	.9967	201	.9973
51	C	C-	168	172	8.073			58	.9855	87	.9905	116	.9921	145	.9932	175	.9950		
52	A	B-	170	180	10.854			56	.975	83	.9875	111	.9932	139	.9953				
"	C	B-	"	"	10.431	42	.976	56	.982	83	.990	119	.9930	139	.9956				
53	C	B-	162	168	11.662	45	.976	60	.9825	89	.9906	121	.9925	149	.9938	179	.9946		
54	B	C	156	166	10.243			60	.980	90	.9875								

54	B	C	156	166	10.141			60	.980	90	.988								
55	B	B	150	168	10.975			59	.983	89	.990	119	.9921						
56	B	B	150	168	11.763			59	.9835	89	.9915	119	.9949	149	.9968	179	.9981		
57	B	B	164	168	12.226	$\Upsilon = 1.0$		59	.980										
58	A	B-	172?	172	12.850			58	.981	87	.9888								
73	D	B+	150	100?	0.559	50	.950	100	.983	150	.9914								
74	D	B-	153	100?	0.575	50	.950	100	.9725	150	.9815	200	.9865						
75	D	A-	156	106?	0.775	46	.955	93	.985	139	.992	185	.9940	231	.9959				
76	D	B-	160	114	0.971	44	.955	88	.984	132	.9917	175	.9951						
78	D	A	168	156	1.184	32	.948	64	.978	96	.9906	128	.9950						
79	D	A	170	162	1.575	31	.947	62	.9765	93	.9860	123	.990	154	.9923				
80	D	A	166	162	1.856	31	.940?	62	.976	93	.989	123	.9943	154	.9964	185	.9976		
81	D	A	169	162	2.022	31	.946	62	.978	93	.9885	123	.993	154	.9957	185	.9977?		
82	D	B	?	162	2.320			62	.973?	93	.9905	123	.9941	154	.9957				
86	E	A	176	180	5.169	28	.963	55	.983	83	.9903	111	.9941	139	.9958				
87	E	A	170	186	6.133			54	.975	81	.991	108	.9955	135	.9971	161	.9980		
88	D	A	169	186	6.345	27	.965	54	.9815	81	.9903	108	.9957	135	.9967	161	.9973		
"	E	A	"	"	6.615	27	.958	54	.9825	81	.9903	108	.9944	135	.9960	161	.9967?		
89	E	A-	164?	180	7.265	28	.956	56	.982	83	.990	111	.9941	139	.9962	167	.9970		
91	D	A	173	177	8.684	28	.962?	57	.9835	85	.9902	113	.9946	141	.9965	169	.9973		
92	E	B	180	174	9.296			58	.9845	86	.9913	115	.9952	144	.9966	173	.9972		
93	E	A-	180	174	9.292			58	.984	86	.9913	115	.9945	144	.9960	173	.9966		
95	E	C+	172	174	9.960			58	.9865	86	.9927	115	.997	144	.9983?				
96	E	C+	170	170	10.266			59	.983	88	.9917	118	.9950	147	.9961				
103	B		117?	126	14.192			79	.9934?	119	.9944	158	.9956	198	.9967	238	.9977		

TABLE 1—Concluded

Shot no.	Rec'd by Detector	Rec. Grade	Shot Depth ft.	Av. Depth water ft.	s.°	γ = 1.5 freq cps	γ = 1.5 V/c₁	γ = 2.0 freq cps	γ = 2.0 V/c₁	γ = 3.0 freq cps	γ = 3.0 V/c₁	γ = 4.0 freq cps	γ = 4.0 V/c₁	γ = 5.0 freq cps	γ = 5.0 V/c₁	γ = 6.0 freq cps	γ = 6.0 V/c₁	γ = 7.0 freq cps	γ = 7.0 V/c₁
105	B	B-	132	126	13.082					119	.9932	158	.9960	198	.9972	238	.9981	278	.9986
106	B	B-	127	121	12.128			83	.984?	123	.9926	165	.9962	207	.9970	247	.9981	289	.9985
107	B	B-	112	125	11.238			80	.9835	120	.9919	160	.9949	200	.9969				
108	B	B-	110	125	10.577			80	.986?	120	.9923	160	.9952	200	.9972				
109	B	B	118	126	9.871					119	.9911	158	.9950						
110	B	B	128	130	9.099			77	.984	115	.9913	154	.9950	192	.9965				
111	B	B	130	140	8.435			72	.982	107	.9906	143	.9941	179	.9970?				
112	B	B	132	138?	8.205			73	.9830?	108	.9917	145	.9946	181	.9965	218	.9980		
113	A	C	125	138	7.662			73	.9855	108	.9920	145	.9955	181	.9972				
"	B	B	"	"	7.485			73	.9840	108	.9923	145	.9953	181	.9968				
114	A	B-	100	138	7.355			72	.9860	108	.9943	145	.9963	181	.9975	218	.9984	Bubble pulse	
"	B	B+	"	"	6.655			72	.9838	108	.9913	145	.9946	181	.9960				
115	A	B-	130	136	6.200			64	.978	96	.9880	128	.9924						
115	B		"	"	6.025			64	.9795	96	.9880	128	.9943	160	.9964	192	.9974		
116	B	B	142	156	5.375			64	.978	96	.9880	128	.9924						
117	A	B	150	156	5.277					96	.9886	128	.9943	160	.9968?			Bubble pulse	
117	B	B+	"	"	4.676	32	.960?	64	.976	96	.9895	128	.9935	160	.9948				
118	B	B-	150	156	4.032			64	.976	96	.9867	128	.9915	160	.9944	192	.9960		
119	A	A	138	155	3.512	32	.943	65	.973	97	.9885	129	.9931	162	.9948				
"	"	"	"	"	3.947			65	.977	97	.989	129	.9934	152	.9965?			Bubble pulse	



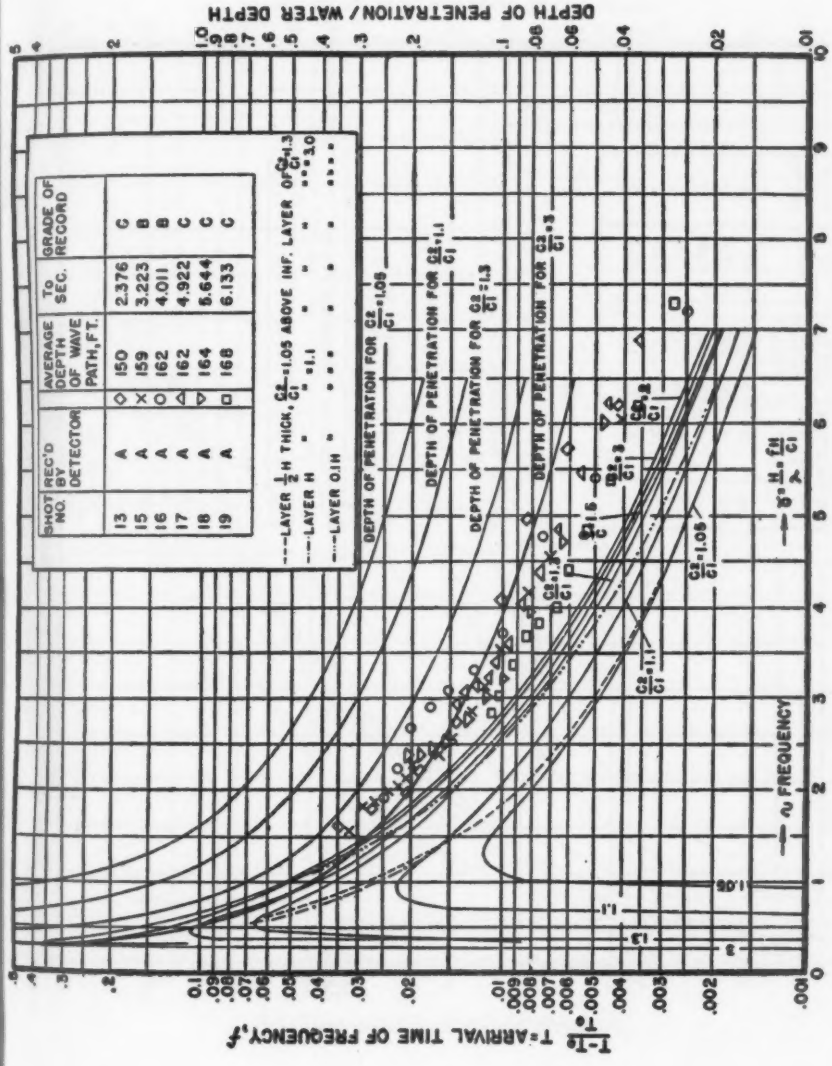

DEPTH OF PENETRATION / WATER DEPTH
T=ARRIVAL TIME OF FREQUENCY, f
SHOT NO.
REC'D BY DETECTOR
AVERAGE DEPTH IN WAVE PATH, FT.
TO SEC.
GRADE OF RECORD
13 A ◇ 150 2.376 C
15 A × 159 3.223 B
16 A ○ 162 4.011 B
17 A △ 162 4.922 C
18 A ▽ 164 5.644 C
19 A □ 168 6.133 C
---LAYER ½ H THICK, C2/C1=1.05 ABOVE INF. LAYER OF C2/C1=1.3
—·—LAYER H " " =1.1 " " " =3.0
——LAYER 0 IN " " =∞ " " " =∞
DEPTH OF PENETRATION FOR C2/C1=1.05
DEPTH OF PENETRATION FOR C2/C1=1.1
DEPTH OF PENETRATION FOR C2/C1=1.3
DEPTH OF PENETRATION FOR C2/C1=3
f FREQUENCY
f = H/λ √(g/2πH)

FIGURE 6.—HIGH-FREQUENCY DISPERSION DATA FROM SHOTS IN AREA I
Data points here fall entirely outside range of group velocity values allowed by Pekeris theory.

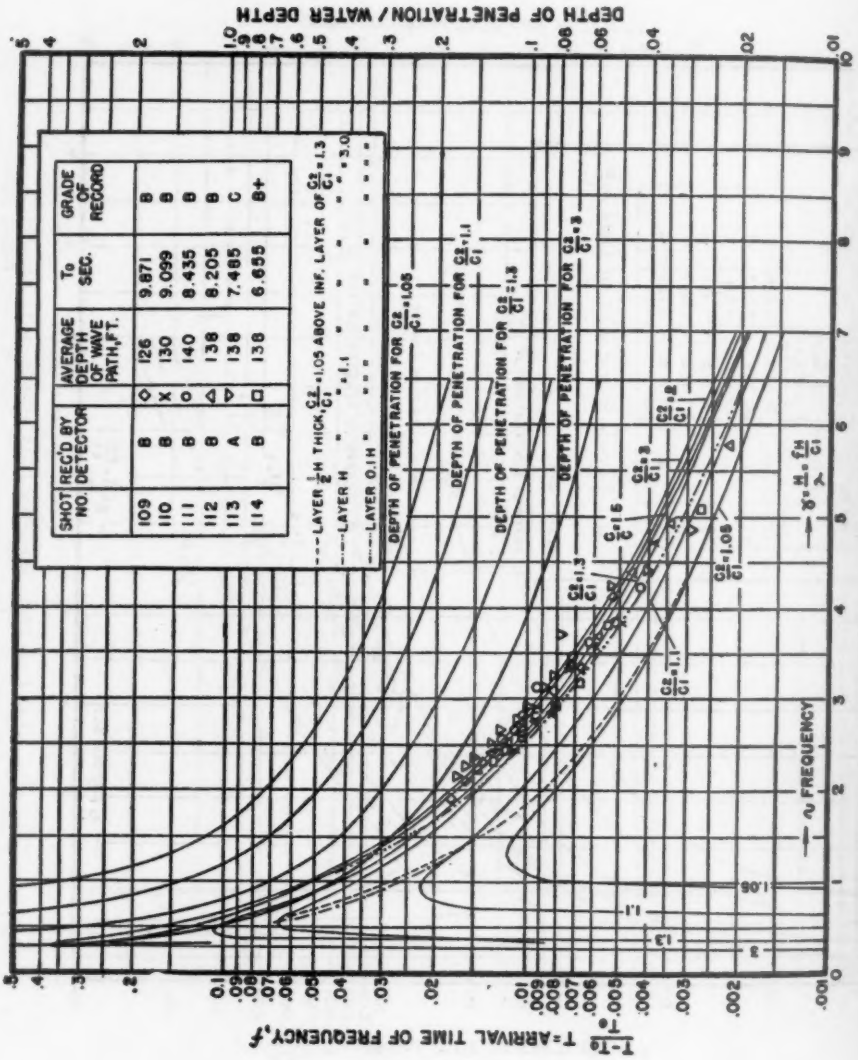

DEPTH OF PENETRATION / WATER DEPTH
T-T0/T0 T=ARRIVAL TIME OF FREQUENCY, f
SHOT NO. | REC'D BY DETECTOR | AVERAGE DEPTH OF WAVE PATH, FT. | To SEC. | GRADE OF RECORD
109 | B | 126 | 9.871 | B
110 | B | 130 | 9.099 | B
111 | B | 140 | 8.435 | B
112 | B | 138 | 8.205 | B
113 | A | 138 | 7.485 | C
114 | B | 138 | 6.655 | B+
LAYER 1/2 H THICK, C2/C1 =1.05 ABOVE INF. LAYER OF C2/C1 =1.3
LAYER H " =1.1 " " " =3.0
LAYER 0.1H " =1.3 " " " =
DEPTH OF PENETRATION FOR C2/C1=1.05
DEPTH OF PENETRATION FOR C2/C1=1.1
DEPTH OF PENETRATION FOR C2/C1=1.3
DEPTH OF PENETRATION FOR C2/C1=3
C2/C1=1.6
C2/C1=3
C2/C1=2
C2/C1=1.3
C2/C1=1.1
C2/C1=1.05
1.05
1.1
1.3
3
f FREQUENCY
θ = H/λ = fH/C1

FIGURE 7.—HIGH-FREQUENCY DISPERSION DATA FROM SHOTS IN AREA IX
Data points indicate energy much higher than maximum allowed by refraction data. (See Figure 4.)

FIGURE 7.—HIGH-FREQUENCY DISPERSION DATA FROM SHOTS IN AREA IX
Data points indicate c_2 speeds much higher than maximum allowed by refraction data. (*See* Figure 4.)

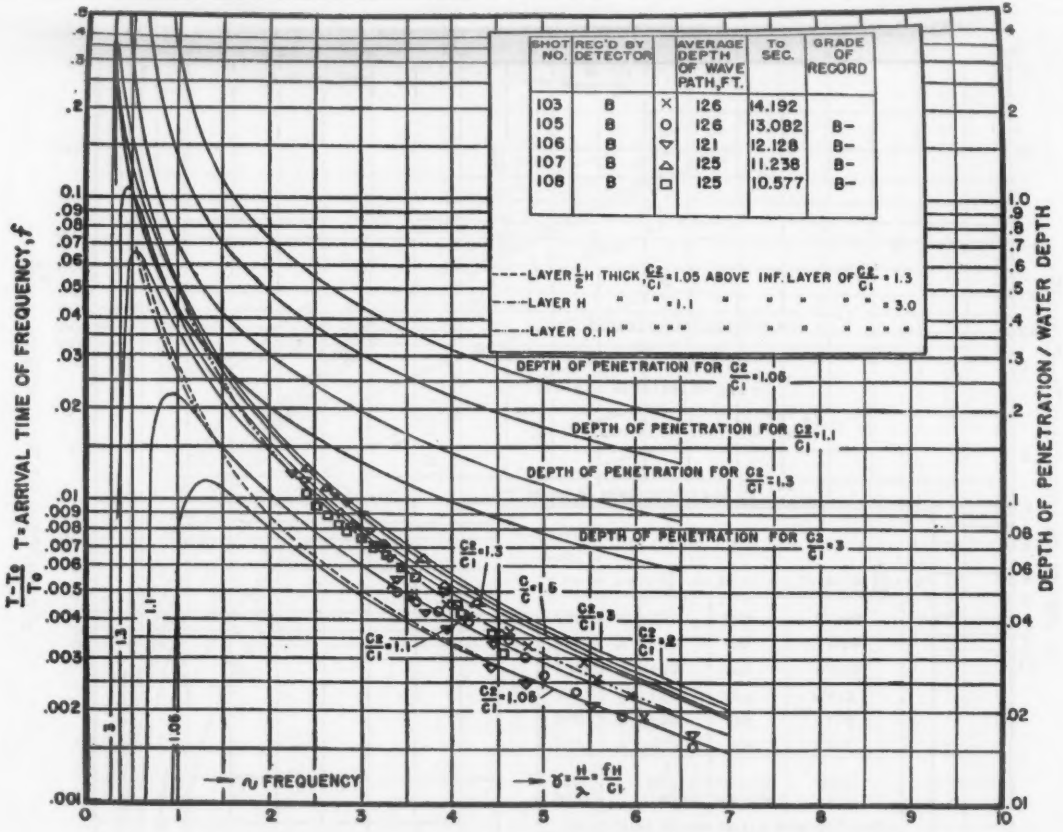

SHOT NO. | REC'D BY DETECTOR | AVERAGE DEPTH OF WAVE PATH, FT. | TO SEC. | GRADE OF RECORD
103 | B | × | 126 | 14.192 |
105 | B | ○ | 126 | 13.082 | B-
106 | B | ▽ | 121 | 12.128 | B-
107 | B | △ | 125 | 11.238 | B-
108 | B | □ | 125 | 10.577 | B-
LAYER ½H THICK, C2/C1 = 1.05 ABOVE INF. LAYER OF C2/C1 = 1.3
LAYER H " " = 1.1 " " " " " = 3.0
LAYER 0.1 H " " " " " " " " "
DEPTH OF PENETRATION FOR C2/C1 = 1.05
DEPTH OF PENETRATION FOR C2/C1 = 1.1
DEPTH OF PENETRATION FOR C2/C1 = 1.3
DEPTH OF PENETRATION FOR C2/C1 = 3
C2/C1 = 1.3
C2/C1 = 1.5
C2/C1 = 3
C2/C1 = 2
C2/C1 = 1.1
C2/C1 = 1.05
∿ FREQUENCY
β = H/λ = fH/C1
(T-T0)/T0 T = ARRIVAL TIME OF FREQUENCY, f
DEPTH OF PENETRATION / WATER DEPTH

FIGURE 8.—HIGH-FREQUENCY DISPERSION DATA FROM SHOTS IN AREA VIII
Data points indicate c_2 speeds in good agreement with maximum allowed by refraction evidence.

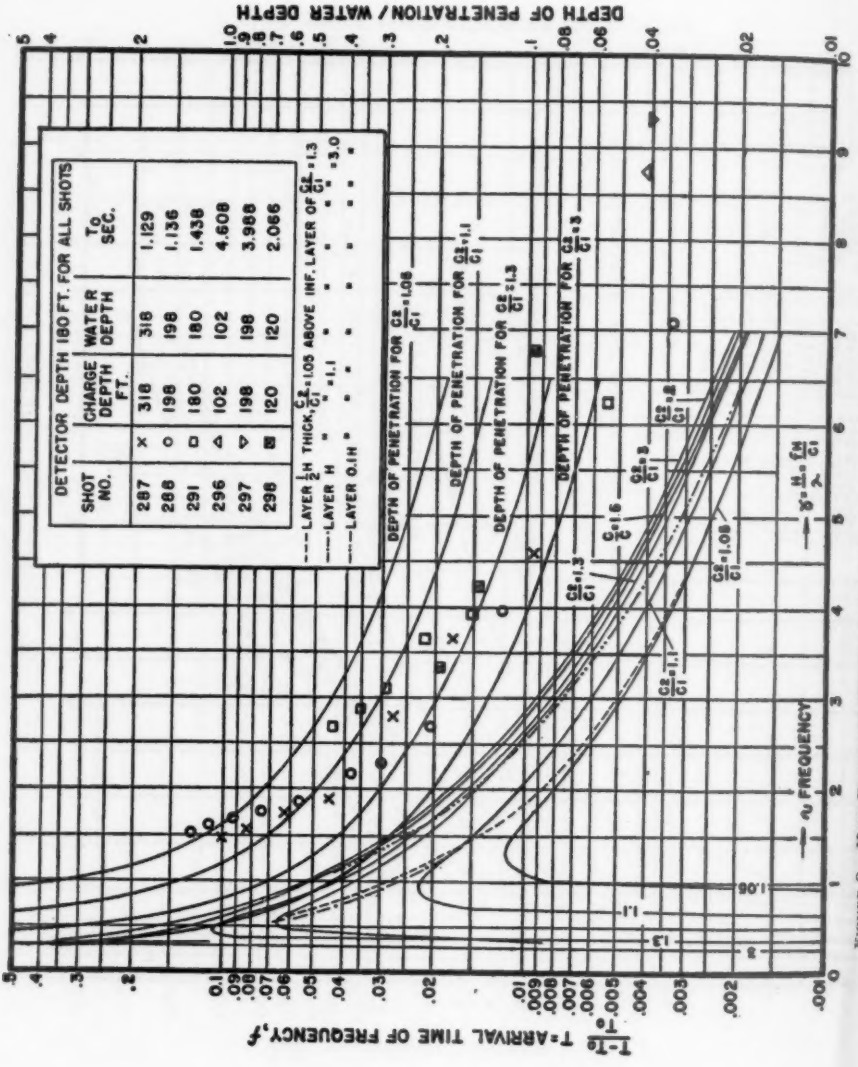

DEPTH OF PENETRATION / WATER DEPTH
T-T0 / T0 = ARRIVAL TIME OF FREQUENCY, f
DETECTOR DEPTH 180 FT. FOR ALL SHOTS
SHOT NO. | CHARGE DEPTH FT. | WATER DEPTH | T0 SEC.
287 × 318 318 1.129
288 ○ 198 198 1.136
291 □ 180 180 1.438
296 △ 102 102 4.608
297 ▽ 198 198 3.988
298 ■ 120 120 2.066
---LAYER 1/2 H THICK, C2/C1 = 1.05 ABOVE INF. LAYER OF C2/C1 = 1.3
——LAYER H " " = 1.1 " " " = 3.0
——LAYER 0.1H " " = " " " = "
DEPTH OF PENETRATION FOR C2/C1 = 1.05
DEPTH OF PENETRATION FOR C2/C1 = 1.1
DEPTH OF PENETRATION FOR C2/C1 = 1.3
DEPTH OF PENETRATION FOR C2/C1 = 3
C2/C1 = 1.6
C2/C1 = 1.3
C2/C1 = 1.1
C2/C1 = 3
C2/C1 = 1.8
C2/C1 = 1.05
1.06
1.1
1.3
3
ν FREQUENCY
γ = H/λ = fH/C1

FIGURE 2.—HIGH-FREQUENCY DISPERSION OBSERVED AT BARBADOS BY EWING AND WORZEL IN 1944
Data from Ewing, Ewing, and Worzel, Geophysics, 12, 1947, 322-329. Points fall on curves of theoretical dispersion computed by Pekeris (1948).

such areas (one of which is represented in Fig. 7) would indicate c_2/c_1 ratios between 2 and 3, which appear much too high on the basis of all prior considerations. In other areas, such as that for Figure 8, the points fall in a more reasonable part of the theoretical region, giving c_2/c_1 values between 1.1 and 1.3 and indicating more likely agreement with theory. It will be shown that the areas giving dispersion patterns of each of the three types have characteristics in common which may give a clue as to the origin of the anomalies observed in the first two kinds of area.

On many of the records the output of two hydrophones as much as several thousand feet apart was recorded simultaneously. Dispersion patterns from some of the shots were often observed on the traces from both hydrophones, but little difference was observed in the respective curves.

Airy Waves

Parts of lagoon where observed.—Pronounced Airy phases were observed on almost all the records made at Enyu from shots fired more than 3 miles away. A few records (Pl. 2) made on the USS GILLIS when anchored off Yuro showed apparent Airy phases so different from those recorded by Worzel and Ewing that their identification as Airy waves is uncertain. Other GILLIS records had no indication of Airy cycles. Many Enyu records of shots fired 13 or more miles from the detectors showed Airy phases much more conspicuous than were the earlier arriving high-frequency water waves. This appears attributable to a greater absorption of the high-frequency energy over long distances and also to the filter settings employed.

Characteristics of the Airy waves as observed on the records.—On the records obtained by Worzel and Ewing the Airy phases were readily distinguishable as the maximum-amplitude cycles at the termination of a train of water waves in which each cycle has a greater amplitude and longer period than the one before. Water-wave arrivals of a number of the Bikini records conformed to this pattern. Record 79 (Pl. 2) is of this type. The majority of the Bikini records, however, show a more complex pattern in which the high-frequency waves appear to die out before the appearance of the lower-frequency wave train leading into the Airy phase.

Records 52 and 53 (Pl. 3, fig. 2) represent an intermediate type of record in which periods increase continuously, although there is a minimum in amplitude just before the onset of the Airy phase. Record 49 (Pl. 3, fig. 1) exemplifies disappearance of the regular water-wave pattern before the emergence of the Airy cycles. Record 60 (Pl. 3, fig. 2) shows a pronounced Airy phase where the earlier higher-frequency waves are scarcely distinguishable from background. Records 78, 85, and 86 (Pl. 3, fig. 1), all conspicuously filtered, show the Airy phase without earlier water waves.

The poor transmission thus indicated for frequencies in the intermediate range between the early water-wave arrivals at 100 cps or more and the much lower-frequency Airy phases is anomalous, but several explanations have been suggested. The intermediate frequencies may be selectively absorbed. No such absorption is called for in this region by Pekeris' normal-mode theory, which for liquid bottoms predicts continuously increasing amplitude as the frequency decreases if the source exhibits exponential pressure fall-off after the instant of explosion. The transmission by earth material of preferred frequencies has long been observed in commercial seismic work but has never been accounted for completely on a theoretical basis. McDermott (1948) as well as Clewell and Simon (1950) have attributed the selective transmission in part to fine laminations within the sedimentary layer. There is some possibility that the amplitude minimum at intermediate frequencies may be instrumental in origin and attributable to resonances or interaction of directly transmitted water pressure with that caused by ground motion below the hydrophone.

Use of Airy waves for studying the bottom.—In many respects the Airy phase gives more information on bottom structure than any other part of the water wave. The penetration of the Airy waves (*see* Figures 6–9) is much greater than that of the higher-frequency water waves. Examination of the group-velocity curves in Figure 2 shows that the curves representing the various c_2/c_1 values are sep-

TABLE 2.—AIRY PHASE DATA

Shot no.	Rec'd by Detector	Shot depth ft.	Av. depth water path	t_0 sec.	Time Airy Phase Arrival, sec.	D_2/D_1 ratio	Period Airy Phase, sec.	Frequency A. Ph., cps	γ	$\frac{\sigma_2}{\sigma_1}$ ratio
20	A	170	172	7.701	8.25	0.933	.050	20	0.69	1.19
21	A	168	160	8.607	9.40	.917	.060?	16.7	.535	1.32
22	A	126	160	9.071	9.60	.945	.050	20	.64	1.23
26	A	168	172	11.089	11.75	.943	.050?	20	.69?	1.19?
28	A	170	172	12.601	13.30	.948	.065?	15.4	.53?	1.35?
29	A	165	175	13.342	14.1	.945	.055	18.2	.63	1.24
32	A	156	175	15.825	16.70	.946	.048	20.9	.73	1.17
34	A	168	175	17.445	18.70	.934	.050	20	.70	1.18
35	A	168	172	18.474	19.70	.937	.050	20	.69	1.19
36	A	156	175	19.710	20.80	.947	.055	18.2	.635	1.24
37	A	162	180	19.177	20.40	.940	.055	18.2	.655	1.23
46	A	166	171	3.297	3.97	.830	.042?	23.8?	.815?	1.13?
"	C	"	"	2.830	3.35	.842	.050	20	.685	1.20
47	C	168	165	3.210	3.35?	.960?	.060?	16.7?	.55?	1.29?
48	A	175	168	4.983	5.40	.925	.050	20	.67	1.21
"	C	"	"	4.523	4.80	.947	.060	16.7	.56	1.30
49	A	168	168	time break bad			.055?	18.2	.62	1.25
"	C	"	"	"	"	"	.060	16.7	.57	1.28
50	A	162	174	6.819	7.50	.912	.055?	18.2?	.63	1.24?
"	C	"	"	6.353	6.70	.950	.060	16.7	.58	1.28
51	A	168	172	8.073	8.50	.950	.060	16.7	.57	1.28
"	C	"	"	7.613	8.20	.930	.055?	18.2?	.63?	1.24?
52	A	170	180	10.854	11.55	.940	.056	17.9	.64	1.23
"	C	"	"	10.431	11.10	.940	.057	17.6	.63	1.24
53	A	162	168	12.125	12.80	.950	.053	19.0	.64	1.23
"	C	"	"	11.662	12.50	.933	.055	18.2	.61	1.25
54	B	156	166	10.141	11.20	.908	.065	15.4	.51	1.38
55	A	150	168	11.150	12.00?	.930?	.050	20	.67	1.21
"	B	"	"	10.975	11.90	.925	.055	18.2	.61	1.25
56	A	150	168	11.930	12.70	.942	.060	16.7	.56	1.30
"	B	"	"	11.763	12.50	.940	.060	16.7	.56	1.30
57	A	164	168	12.390	13.20	.939	.050	20	.675	1.20
"	B	"	"	12.225	13.00	.942	.055	18.2	.61	1.25
58	A	166	172	13.100	14.10?	.930?	.050	20	.69	1.19
"	B	"	"	12.920	14.00	.924	.050	20	.69	1.19
59	A	170	172	13.550	14.30	.948	.055	18.2	.63	1.24
"	B	"	"	13.393	14.10	.950	.053	18.9	.65	1.23
60	B	168	175	13.697	14.50	.946	.060	16.7	.58	1.27

TABLE 2—Concluded

Shot no.	Rec'd by Detector	Shot depth ft.	Av. depth wave path ft.	T_2 sec.	Time Airy Phase Arrival, sec.	$\frac{U}{c_1}$ ratio	Period Airy Phase, sec.	Frequency A. Ph., cps	γ	$\frac{c_2}{c_1}$ ratio
61	B	168	175	14.517	15.20	.955	.060	16.7	.58	1.27
62	B	170	175	14.526	15.50	.938	.055	18.2	.63	1.24
63	B	172	175	14.982	16.00	.937	.057	17.5	.615	1.25
64	B	165	170	15.240	16.30	.936	.060	16.7	.57	1.29
78	B	168	156	16.825	17.80	.945	.060	16.7	.52	1.37
79	B	170	162	16.450?	17.90	.920?	.060	16.7	.54	1.34
80	B	166	162	17.030	18.70	.913	.055	18.2	.59	1.27
81	B	169	162	17.035	18.90	.903	.065	15.4	.50	1.40
83	B	168	170	16.460	17.40	.946	.057	17.6	.60	1.26
84	B	172	168	16.380	17.30	.948	.057	17.6	.59	1.27
85	B	170	172	16.522	17.40	.950	.050	20	.68	1.20
86	B	176	180	16.537	17.60	.940	.053	18.9	.68	1.20
87	B	170	186	16.705	17.70	.945	.053	18.9	.70	1.18
88	B	169	186	16.827	18.01	.935	.060	16.7	.62	1.25
89	B	164	180	17.120	18.50	.928	.057	17.5	.63	1.24
90	B	170	167	17.455	19.0	.918	.050	20	.67	1.21
91	B	173	177	17.610	18.50	.952	.057	17.6	.61	1.25
93	B	180	174	17.830	19.10	.935	.057	17.6	.63	1.24
					Av.	.935			Av.	1.25
111	B	130	140	8.435	8.80	.955	.060	16.7	.47	1.43
112	B	132	138	8.205	8.50	.965	.055?	18.2?	.51?	1.39?
113	B	125	138	7.485	7.80	.960	.065	15.4	.43	1.60
114	B	100	138	6.655	6.90	.963	.065	15.4	.43	1.60
115	B	130	156	6.025	6.30	.955	.065?	15.4?	.48?	1.44?
116	B	142	156	5.375	5.60	.950	.060	16.7	.52	1.37
117	B	150	156	4.676	5.30	.885	.060	16.7	.52	1.37
					Av.	.947			Av.	1.45

arated by the greatest amount in the region of minimum-group velocity. Thus, for the two-layer case, the Airy data make possible much greater resolution in interpolating between the c_2/c_1 curves than do data on the higher-frequency dispersion. Experimental uncertainties in this frequency region are generally much less critical than at higher frequencies.

Velocity-frequency relationships.—Table 2 summarizes the data on frequency and group velocity obtained from all identifiable Airy waves. The records (Nos. 111–117) from the Enyu-Cherry profile along the southern edge of the lagoon are grouped separately from the others, since the Airy phases seem to show small but consistent differences from those observed on the other records.

The precision of the Airy group velocities

and the frequencies is limited by several sources of uncertainty. Often the exact time of the highest Airy peak could not be estimated to closer than the nearest 0.1 sec. In many cases the period of the Airy waves could be determined only to the nearest .010 second owing to distortion of the wave by natural and instrumental background. Finally, the uncertainty in the proper value to use for H in determining γ for the shots over irregular bottoms also could introduce substantial error.

For the records made at Enyu of shots along all profiles but the Enyu-Cherry, the Airy phases indicate an average velocity of sound in the lagoon bottom 1.25 times the velocity of sound in water. This speed represents the average within a layer extending about 350 feet below the water bottom. It is so weighted, however, that the velocity in the upper part of the section contributes somewhat more substantially. The refraction data indicate that the ratio of speeds should average no higher than 1.30 for a layer extending approximately 400 feet below the water bottom. This average, as pointed out previously, gives more weight to the speed in the lower part of the section. Since compaction should cause an increase of speed in the first few hundred feet, one might expect the speed determined by refraction methods to be somewhat greater than that determined from the Airy data. The small difference between the two is thus in the direction one would predict.

The anomalously high speed indicated for the uppermost section along the central part of the Enyu-Cherry profile cannot be checked by reference to the refraction data since there is no refraction control on the speed in the uppermost zone within many miles of this area.

Multiple Airy phases.—On 20 records the Airy phases were characterized by series of 2 to 6 maxima, all of the same frequency; adjacent pairs were 0.3 to 0.4 sec. apart. Records 78 and 79 (Pl. 3, fig. 1) and all five traces shown in Figure 2 of Plate 3 illustrate this feature.

Almost certainly, all the Airy maxima after the first represent bubble pulses, even though the high-frequency bursts usually associated with such pulses are missing on the Mk 1 traces. Some of the records showing multiple Airy phases have high-frequency traces (from a crystal hydrophone) indicating weak secondary bursts characteristic of bubble pulsing several tenths of a second after the first water-wave arrival. Intervals between successive bursts were compared with those between corresponding Airy maxima on the same records. Results were as follows:

Record No.	Bubble pulse interval, sec. (high frequency)	Interval between Airy maxima, sec.
21	0.42?	0.35?
46	0.38	0.38
47	0.38	0.42
49	0.41	0.41
51	0.41	0.41
59	0.50	0.47

This is convincing evidence that the subsequent Airy maxima are associated with bubble pulses. It appears that the energy of the pulses on these records is much more strongly concentrated at lower frequencies than in Worzel and Ewing's records. The bubble pulses from shots on the Enyu-Cherry profile exhibit water waves much richer in high frequencies than those from other portions of the lagoon.

Summary of Observed Dispersion Data

The experimental observations presented thus far may be summarized as follows:

(1) Application of the Pekeris normal-mode theory to the higher-frequency dispersion data show that most of the shots in the deep part of the lagoon give dispersion curves lying entirely outside the range of values permitted by the theory. All group velocities from these shots are too low.

(2) Many shots fired in water between 25 and 30 fathoms and nearer the edge of the lagoon yield dispersion data within the theoretically possible range, but the group velocities are still lower than those which theory would require if the bottom velocities are below the limits allowed by refraction data.

(3) In general, shots in the shallow water near the lagoon's edge give high-frequency dispersion data indicating, on the basis of normal-mode theory, bottom velocities quite

compatible with the refraction data. Such records indicate an increase of speed with depth as follows:

Depth d (ft.)	22	30	40 and over
c_4/c_1 (aver. above d)	1.05	1.1	1.3

(4) Airy phases on the records give information on the velocity of the bottom material down to about twice the water depth, as follows:

(a) Data from the Enyu-Namu, Enyu-Yuro, and Yuro-Cherry profiles indicated a quite consistent *average* bottom velocity 1.25 times that in water. This agrees well with refraction evidence that the uppermost 400 feet should have an average speed ratio of no more than 1.30.

(b) Along the west-central part of the Enyu-Cherry profile, the Airy data indicate a somewhat higher velocity, approximately 1.45 times that in water. The precision of the data in this region is much poorer than that of the other Airy data.

Interpretation of Results

General Statement

The dispersion data obtained at Bikini can be explained by the Pekeris theory for two liquid layers only for the Airy phases and for most of the higher-frequency water waves from explosions in relatively shallow water near the edge of the lagoon. For the remainder of the shots, the high-frequency dispersion data do not generally conform to this theory; the group velocities are lower for the frequency, water depth, and maximum bottom speed (as ascertained from refraction) than the theory predicts. The divergences between theory and observation are too large and too consistent to be accounted for by known experimental uncertainties.

Comparison of Bikini Anomalies with those at Barbados and Orinoco Delta

The seismic work of Worzel and Ewing (1948) at Barbados and in the Orinoco delta yielded dispersion curves with anomalously low group velocities, similar in many ways to the typical Bikini records from deep-water shots. The Barbados anomalies are so similar to those from Bikini that any explanation of one might well cast light on the other (Fig. 9).

Worzel and Ewing (1948) have pointed out that the bottom here has considerable slope transverse to the wave trajectory and is also quite irregular in depth. In the Orinoco delta, where the same anomaly is observed, the bottom is known to be smooth and to dip no more than 20 ft/mi. From their refraction data, they found that at Barbados the velocity of the bottom is no more than 1.12 times that of water down to 2700 feet, while at Orinoco it is slightly less than 1.10 times water velocity down to 1800 feet. Senn (1946) has published a report on the subsurface geology of Barbados as ascertained from its water wells, some of which are more than 300 feet deep. He deduces that the uppermost formation of the submerged western coastal shelf is a 100-foot thick moderately permeable Pleistocene limestone reef formation called the Coral-rock. Beneath the water of the north coastal shelf, it appears to have been completely eroded away. Below the Coral-rock is the "Oceanic formation", a fine-grained water saturated sediment consisting of abundant *globigerina* marls and radiolarian marls and clays. The two formations have sufficiently different physical properties that one would expect their seismic speeds to be somewhat different, and this may give a clue to the anomalous dispersion observed along the western coastal shelf.

Possible Explanations for Anomalous Dispersion

General statement.—Almost any departure from the assumption of a homogeneous semi-infinite "liquid" bottom material below a flat interface might conceivably explain the discrepancy between theory and observation. Consider the following proposed explanations:

(1) Departure from uniform water depth.

(2) Solid rather than liquid bottom.

(3) Two or more layers below water bottom rather than a single layer (as in the three-layer case considered by Pekeris).

(4) Bottom velocity lower that of water.

(5) Horizontal change in elastic characteristics of the bottom.

Nonuniform water depth.—The normal-mode theory of Pekeris assumes that all interfaces are horizontal. Any relief of the water bottom might thus give rise to a discrepancy between theory and observation. Several lines of evidence, however, indicate that such an effect should not be significant as long as the difference in depth is no more than about half the water depth. Pekeris found he could apply his theory successfully to areas where the shots were 50 per cent deeper than the hydrophones at which they were received. The dispersion patterns recorded from the same shot by detectors at Bikini in 64 and 134 feet of water, respectively, showed almost no difference in their characteristics at the two detectors. Anomalously low group velocities were observed moreover at Orinoco, where the bottom has very little slope.

Solid bottom.—The theory with which the present data are compared assumes a liquid bottom. Some question may arise as to whether the observed discrepancies could be caused by a bottom that transmits transverse waves as well as longitudinal ones, thus behaving as a solid. Press and Ewing (1948a) have extended the Pekeris theory to cover the dispersion of water waves over a solid bottom. The curve they obtain for group velocity vs. frequency consists of two branches, each corresponding to a different mode. One resembles the curve for a liquid bottom with the same speed as that of longitudinal waves in the solid. It also has a group-velocity minimum almost coincident with that for the corresponding liquid case. The other branch has a group-velocity minimum at about three times the frequency of the other minimum and in addition a maximum located between the two minima. The Bikini records give no evidence for additional group-velocity turning points at frequencies several times as high as the one for which the liquid-bottom theory gives results consistent with the refraction data. Finally, there is no geologic or refraction evidence for any hard rock within the first few hundred feet of the lagoon bottom.

Vertical changes in velocity.—If the material at the bottom of Bikini lagoon consists of unconsolidated sands and algal debris, compaction will cause the velocity of sound in the material to increase with depth below the water bottom. In this event, the two-layer theory which assumes an entirely homogeneous bottom would, of course, not be valid. The dispersion curve for this case, however, can be readily predicted with reference to the conventional two-layer curves. Its trajectory on the $(T - T_0)/T_0$ vs. γ plot will cross the standard curves between the limits represented by the highest and lowest velocities in the changing section. Pekeris has worked out the curves exactly for a number of cases where there are three distinct layers, each of constant speed. Some of these are plotted as broken lines along with the various two-layer curves used for reference on the dispersion plots of Figures 6–9. Since all the three-layer curves lie within the range which theory allows for the two-layer curves, no velocity variation dependent on depth alone can explain the existence of group velocities outside this range.

Low-velocity bottom.—Press and Ewing (1948b) have applied the normal-mode theory to the case where the bottom material has a lower speed than that of water. They show that such a configuration, with a low-velocity layer between two higher-velocity layers, should yield a dispersion curve having two branches. Along one of these, the high frequencies travel with the speed of sound in the low-velocity layer. Along the other branch, there is considerable attenuation of all frequencies except the highest, which arrive with the speed of sound in water. The pattern of high-frequency arrivals traveling at a lower speed than that of water has never been observed. High-frequency, nondispersive arrivals traveling with the water speed were obtained from all Bikini shots recorded with the detectors off the USS GILLIS when anchored near Cherry Island. These were shots no. 110 to 120 along the eastern part of the Enyu-Cherry profile. Since the same shots when received at Enyu give the usual dispersion pattern, there is probably a low-velocity bottom only under the western end of the trajectory.

Horizontal lithology change.—The normal-mode theory as applied to the propagation of explosion waves through the water has been developed on the assumption that the elastic properties of the bottom vary only with depth over the entire wave trajectory. Any horizontal

change in these properties would introduce new boundary conditions which could change the dispersion characteristics markedly and render the theory in its present form inapplicable. Possibly the theory can be extended to cover rigorously the introduction of a vertical discontinuity in speeds as a new boundary condition.

According to Emery's preliminary report (1948) on the bottom composition of Bikini lagoon, several zones of different bottom lithologies are concentric with the lagoon's boundary. The outermost ring covering the shallow terrace at the edge is coarse sand which grades lagoonward into fine sand. At 15 to 20 fathoms depth a disc of *Halimeda* debris begins, and this covers the entire central portion of the lagoon. No specific lithologic boundaries have yet been mapped, but the gradation from finer to coarser bottom sediments as one goes from the center to the periphery of the lagoon appears well established. Since all recording hydrophones were within 3000 feet of shore, it is expected that the receiving end of all wave trajectories would be in the relatively coarse material characteristic of the lagoon's edge. Where the explosion was in the deeper part of the lagoon, this end of the trajectory would have been over the fine sediments characteristic of the center. For these shots one would therefore expect some kind of horizontal lithologic change beneath the path of the water waves. Shots near the edge of the lagoon, on the other hand, would then have both the beginning and end of their trajectories over the same kind of bottom material, and one would expect the results to conform more closely to a theory assuming lateral homogeneity.

As a working hypothesis, let us assume that the discrepancies between the observed high-frequency dispersion and the theoretical predictions can be correlated with lithologic changes (down to about 100 feet below the bottom) between the two ends of the wave trajectory. It appears likely on the basis of available evidence that the bottom constitution is related to the depth of water and the distance from the lagoon's edge. It might be instructive to tabulate the high-frequency dispersion data with respect to these two variables. The data are separated according to shot location; the areas specified on the map (Pl. 1) by the Roman numerals are considered individually. In the following tabulation the agreement of the dispersion data with the refraction data is designated by one of three letters. *A* indicates close agreement between refraction and dispersion, (Fig. 6); *B* indicates poor agreement, but the difference is not so great that it could not conceivably be explained by experimental uncertainty (Fig. 7). *C* indicates that the observed dispersion is definitely outside the region which the normal-mode theory allows (Fig. 8).

Area	Recorded off Enyu (E) or Yuro (Y)	Average depth (fathoms)	Aver. distance from edge (naut. miles)	Agreement of dispersion data and refraction data
I	E	29	3.2	C
I	E	31	4.0	C
III	E	28	2.2	B
IV	Y	22	1.2	A
V	Y	29	2.0	C
VI	Y	30	4.5	B
VII	Y	27	2.7	A
VIII	E	22	1.1	A
IX	E	23	1.2	B
X	E	27	1.2	C

Thus better agreement is observed where the shots are in shallow water and near the edge. The principal exceptions are in areas VI and VII, which are in deep enough water and are close enough to the center of the lagoon that, on the basis of the results in other areas, one would not expect as good agreement as is obtained. In general the table supports the hypothesis that the better agreement is obtained where there is the least change in water depth and distance from the lagoon's edge between shot and detector.

The high-frequency dispersion data are applicable to depths of the order of 100 to 150 feet, while the Airy-phase data represent penetration to about 400 feet. The Airy dispersion exhibits no such variation with shot position but indicates bottom velocities that are uniform everywhere in the lagoon except along the southern edge. These velocities give good agreement with those that would be expected from the refraction data. Thus any lithologic change necessary to explain the discrepancy in the

TABLE 2.—AIRY PHASE DATA

Shot no.	Rec'd by Detector	Shot depth ft.	Av. depth water ft.	t_o sec.	Time Airy Phase Arrival, sec.	D/d ratio	Period Airy Phase, sec.	Frequency A. Ph., cps	γ	σ/σ' ratio
20	A	170	172	7.701	8.25	0.933	.050	20	0.69	1.19
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22	A	126	160	9.071	9.60	.945	.050	20	.64	1.23
26	A	168	172	11.089	11.75	.943	.050?	20	.69?	1.19?
28	A	170	172	12.601	13.30	.948	.065?	15.4	.53?	1.35?
29	A	165	175	13.342	14.1	.945	.055	18.2	.63	1.24
32	A	156	175	15.825	16.70	.946	.048	20.9	.73	1.17
34	A	168	175	17.445	18.70	.934	.050	20	.70	1.18
35	A	168	172	18.474	19.70	.937	.050	20	.69	1.19
36	A	156	175	19.710	20.80	.947	.055	18.2	.635	1.24
37	A	162	180	19.177	20.40	.940	.055	18.2	.655	1.23
46	A	166	171	3.297	3.97	.830	.042?	23.8?	.815?	1.13?
"	C	"	"	2.830	3.35	.842	.050	20	.685	1.20
47	C	168	165	3.210	3.35?	.960?	.060?	16.7?	.55?	1.29?
48	A	175	168	4.983	5.40	.925	.050	20	.67	1.21
"	C	"	"	4.523	4.80	.947	.060	16.7	.56	1.30
49	A	168	168	time break bad			.055?	18.2	.62	1.25
"	C	"	"	" " "			.060	16.7	.57	1.28
50	A	162	174	6.819	7.50	.912	.055?	18.2?	.63	1.24?
"	C	"	"	6.353	6.70	.950	.060	16.7	.58	1.28
51	A	168	172	8.073	8.50	.950	.060	16.7	.57	1.28
"	C	"	"	7.613	8.20	.930	.055?	18.2?	.63?	1.24?
52	A	170	180	10.854	11.55	.940	.056	17.9	.64	1.23
"	C	"	"	10.431	11.10	.940	.057	17.6	.63	1.24
53	A	162	168	12.125	12.80	.950	.053	19.0	.64	1.23
"	C	"	"	11.662	12.50	.933	.055	18.2	.61	1.25
54	B	156	166	10.141	11.20	.908	.065	15.4	.51	1.38
55	A	150	168	11.150	12.00?	.930?	.050	20	.67	1.21
"	B	"	"	10.975	11.90	.925	.055	18.2	.61	1.25
56	A	150	168	11.930	12.70	.942	.060	16.7	.56	1.30
"	B	"	"	11.763	12.50	.940	.060	16.7	.56	1.30
57	A	164	168	12.390	13.20	.939	.050	20	.675	1.20
"	B	"	"	12.225	13.00	.942	.055	18.2	.61	1.25
58	A	166	172	13.100	14.10?	.930?	.050	20	.69	1.19
"	B	"	"	12.920	14.00	.924	.050	20	.69	1.19
59	A	170	172	13.550	14.30	.948	.055	18.2	.63	1.24
"	B	"	"	13.393	14.10	.950	.053	18.9	.65	1.23
60	B	168	175	13.697	14.50	.946	.060	16.7	.58	1.27

TABLE 2—Concluded

Shot no.	Rec'd by Detector	Shot depth ft.	Av. depth wave path	T_0 sec.	Time Airy Phase Arrival, sec.	$\frac{U}{c_1}$ ratio	Period Airy Phase, sec.	Frequency A. Ph., cps	γ	$\frac{c_2}{c_1}$ ratio
61	B	168	175	14.517	15.20	.955	.060	16.7	.58	1.27
62	B	170	175	14.526	15.50	.938	.055	18.2	.63	1.24
63	B	172	175	14.982	16.00	.937	.057	17.5	.615	1.25
64	B	165	170	15.240	16.30	.936	.060	16.7	.57	1.29
78	B	168	156	16.825	17.80	.945	.060	16.7	.52	1.37
79	B	170	162	16.450?	17.90	.920?	.060	16.7	.54	1.34
80	B	166	162	17.030	18.70	.913	.055	18.2	.59	1.27
81	B	169	162	17.035	18.90	.903	.065	15.4	.50	1.40
83	B	168	170	16.460	17.40	.946	.057	17.6	.60	1.26
84	B	172	168	16.380	17.30	.948	.057	17.6	.59	1.27
85	B	170	172	16.522	17.40	.950	.050	20	.68	1.20
86	B	176	180	16.537	17.60	.940	.053	18.9	.68	1.20
87	B	170	186	16.705	17.70	.945	.053	18.9	.70	1.18
88	B	169	186	16.827	18.01	.935	.060	16.7	.62	1.25
89	B	164	180	17.120	18.50	.928	.057	17.5	.63	1.24
90	B	170	167	17.455	19.0	.918	.050	20	.67	1.21
91	B	173	177	17.610	18.50	.952	.057	17.6	.61	1.25
93	B	180	174	17.830	19.10	.935	.057	17.6	.63	1.24
					Av.	.935			Av.	1.25
111	B	130	140	8.435	8.80	.955	.060	16.7	.47	1.43
112	B	132	138	8.205	8.50	.965	.055?	18.2?	.51?	1.39?
113	B	125	138	7.485	7.80	.960	.065	15.4	.43	1.60
114	B	100	138	6.655	6.90	.963	.065	15.4	.43	1.60
115	B	130	156	6.025	6.30	.955	.065?	15.4?	.48?	1.44?
116	B	142	156	5.375	5.60	.950	.060	16.7	.52	1.37
117	B	150	156	4.676	5.30	.885	.060	16.7	.52	1.37
					Av.	.947			Av.	1.45

arated by the greatest amount in the region of minimum-group velocity. Thus, for the two-layer case, the Airy data make possible much greater resolution in interpolating between the c_2/c_1 curves than do data on the higher-frequency dispersion. Experimental uncertainties in this frequency region are generally much less critical than at higher frequencies.

Velocity-frequency relationships.—Table 2 summarizes the data on frequency and group velocity obtained from all identifiable Airy waves. The records (Nos. 111–117) from the Enyu-Cherry profile along the southern edge of the lagoon are grouped separately from the others, since the Airy phases seem to show small but consistent differences from those observed on the other records.

The precision of the Airy group velocities

and the frequencies is limited by several sources of uncertainty. Often the exact time of the highest Airy peak could not be estimated to closer than the nearest 0.1 sec. In many cases the period of the Airy waves could be determined only to the nearest .010 second owing to distortion of the wave by natural and instrumental background. Finally, the uncertainty in the proper value to use for H in determining γ for the shots over irregular bottoms also could introduce substantial error.

For the records made at Enyu of shots along all profiles but the Enyu-Cherry, the Airy phases indicate an average velocity of sound in the lagoon bottom 1.25 times the velocity of sound in water. This speed represents the average within a layer extending about 350 feet below the water bottom. It is so weighted, however, that the velocity in the upper part of the section contributes somewhat more substantially. The refraction data indicate that the ratio of speeds should average no higher than 1.30 for a layer extending approximately 400 feet below the water bottom. This average, as pointed out previously, gives more weight to the speed in the lower part of the section. Since compaction should cause an increase of speed in the first few hundred feet, one might expect the speed determined by refraction methods to be somewhat greater than that determined from the Airy data. The small difference between the two is thus in the direction one would predict.

The anomalously high speed indicated for the uppermost section along the central part of the Enyu-Cherry profile cannot be checked by reference to the refraction data since there is no refraction control on the speed in the uppermost zone within many miles of this area.

Multiple Airy phases.—On 20 records the Airy phases were characterized by series of 2 to 6 maxima, all of the same frequency; adjacent pairs were 0.3 to 0.4 sec. apart. Records 78 and 79 (Pl. 3, fig. 1) and all five traces shown in Figure 2 of Plate 3 illustrate this feature.

Almost certainly, all the Airy maxima after the first represent bubble pulses, even though the high-frequency bursts usually associated with such pulses are missing on the Mk 1 traces. Some of the records showing multiple Airy phases have high-frequency traces (from a crystal hydrophone) indicating weak secondary bursts characteristic of bubble pulsing several tenths of a second after the first water-wave arrival. Intervals between successive bursts were compared with those between corresponding Airy maxima on the same records. Results were as follows:

Record No.	Bubble pulse interval, sec. (high frequency)	Interval between Airy maxima, sec.
21	0.42?	0.35?
46	0.38	0.38
47	0.38	0.42
49	0.41	0.41
51	0.41	0.41
59	0.50	0.47

This is convincing evidence that the subsequent Airy maxima are associated with bubble pulses. It appears that the energy of the pulses on these records is much more strongly concentrated at lower frequencies than in Worzel and Ewing's records. The bubble pulses from shots on the Enyu-Cherry profile exhibit water waves much richer in high frequencies than those from other portions of the lagoon.

Summary of Observed Dispersion Data

The experimental observations presented thus far may be summarized as follows:

(1) Application of the Pekeris normal-mode theory to the higher-frequency dispersion data show that most of the shots in the deep part of the lagoon give dispersion curves lying entirely outside the range of values permitted by the theory. All group velocities from these shots are too low.

(2) Many shots fired in water between 25 and 30 fathoms and nearer the edge of the lagoon yield dispersion data within the theoretically possible range, but the group velocities are still lower than those which theory would require if the bottom velocities are below the limits allowed by refraction data.

(3) In general, shots in the shallow water near the lagoon's edge give high-frequency dispersion data indicating, on the basis of normal-mode theory, bottom velocities quite

compatible with the refraction data. Such records indicate an increase of speed with depth as follows:

Depth d (ft.)	22	30	40 and over
c_4/c_1 (aver. above d)	1.05	1.1	1.3

(4) Airy phases on the records give information on the velocity of the bottom material down to about twice the water depth, as follows:

(a) Data from the Enyu-Namu, Enyu-Yuro, and Yuro-Cherry profiles indicated a quite consistent *average* bottom velocity 1.25 times that in water. This agrees well with refraction evidence that the uppermost 400 feet should have an average speed ratio of no more than 1.30.

(b) Along the west-central part of the Enyu-Cherry profile, the Airy data indicate a somewhat higher velocity, approximately 1.45 times that in water. The precision of the data in this region is much poorer than that of the other Airy data.

Interpretation of Results

General Statement

The dispersion data obtained at Bikini can be explained by the Pekeris theory for two liquid layers only for the Airy phases and for most of the higher-frequency water waves from explosions in relatively shallow water near the edge of the lagoon. For the remainder of the shots, the high-frequency dispersion data do not generally conform to this theory; the group velocities are lower for the frequency, water depth, and maximum bottom speed (as ascertained from refraction) than the theory predicts. The divergences between theory and observation are too large and too consistent to be accounted for by known experimental uncertainties.

Comparison of Bikini Anomalies with those at Barbados and Orinoco Delta

The seismic work of Worzel and Ewing (1948) at Barbados and in the Orinoco delta yielded dispersion curves with anomalously low group velocities, similar in many ways to the typical Bikini records from deep-water shots. The Barbados anomalies are so similar to those from Bikini that any explanation of one might well cast light on the other (Fig. 9).

Worzel and Ewing (1948) have pointed out that the bottom here has considerable slope transverse to the wave trajectory and is also quite irregular in depth. In the Orinoco delta, where the same anomaly is observed, the bottom is known to be smooth and to dip no more than 20 ft/mi. From their refraction data, they found that at Barbados the velocity of the bottom is no more than 1.12 times that of water down to 2700 feet, while at Orinoco it is slightly less than 1.10 times water velocity down to 1800 feet. Senn (1946) has published a report on the subsurface geology of Barbados as ascertained from its water wells, some of which are more than 300 feet deep. He deduces that the uppermost formation of the submerged western coastal shelf is a 100-foot thick moderately permeable Pleistocene limestone reef formation called the Coral-rock. Beneath the water of the north coastal shelf, it appears to have been completely eroded away. Below the Coral-rock is the "Oceanic formation", a fine-grained water saturated sediment consisting of abundant *globigerina* marls and radiolarian marls and clays. The two formations have sufficiently different physical properties that one would expect their seismic speeds to be somewhat different, and this may give a clue to the anomalous dispersion observed along the western coastal shelf.

Possible Explanations for Anomalous Dispersion

General statement.—Almost any departure from the assumption of a homogeneous semi-infinite "liquid" bottom material below a flat interface might conceivably explain the discrepancy between theory and observation. Consider the following proposed explanations:

(1) Departure from uniform water depth.

(2) Solid rather than liquid bottom.

(3) Two or more layers below water bottom rather than a single layer (as in the three-layer case considered by Pekeris).

(4) Bottom velocity lower that of water.

(5) Horizontal change in elastic characteristics of the bottom.

Nonuniform water depth.—The normal-mode theory of Pekeris assumes that all interfaces are horizontal. Any relief of the water bottom might thus give rise to a discrepancy between theory and observation. Several lines of evidence, however, indicate that such an effect should not be significant as long as the difference in depth is no more than about half the water depth. Pekeris found he could apply his theory successfully to areas where the shots were 50 per cent deeper than the hydrophones at which they were received. The dispersion patterns recorded from the same shot by detectors at Bikini in 64 and 134 feet of water, respectively, showed almost no difference in their characteristics at the two detectors. Anomalously low group velocities were observed moreover at Orinoco, where the bottom has very little slope.

Solid bottom.—The theory with which the present data are compared assumes a liquid bottom. Some question may arise as to whether the observed discrepancies could be caused by a bottom that transmits transverse waves as well as longitudinal ones, thus behaving as a solid. Press and Ewing (1948a) have extended the Pekeris theory to cover the dispersion of water waves over a solid bottom. The curve they obtain for group velocity vs. frequency consists of two branches, each corresponding to a different mode. One resembles the curve for a liquid bottom with the same speed as that of longitudinal waves in the solid. It also has a group-velocity minimum almost coincident with that for the corresponding liquid case. The other branch has a group-velocity minimum at about three times the frequency of the other minimum and in addition a maximum located between the two minima. The Bikini records give no evidence for additional group-velocity turning points at frequencies several times as high as the one for which the liquid-bottom theory gives results consistent with the refraction data. Finally, there is no geologic or refraction evidence for any hard rock within the first few hundred feet of the lagoon bottom.

Vertical changes in velocity.—If the material at the bottom of Bikini lagoon consists of unconsolidated sands and algal debris, compaction will cause the velocity of sound in the material to increase with depth below the water bottom. In this event, the two-layer theory which assumes an entirely homogeneous bottom would, of course, not be valid. The dispersion curve for this case, however, can be readily predicted with reference to the conventional two-layer curves. Its trajectory on the $(T - T_s)/T_s$ vs. γ plot will cross the standard curves between the limits represented by the highest and lowest velocities in the changing section. Pekeris has worked out the curves exactly for a number of cases where there are three distinct layers, each of constant speed. Some of these are plotted as broken lines along with the various two-layer curves used for reference on the dispersion plots of Figures 6–9. Since all the three-layer curves lie within the range which theory allows for the two-layer curves, no velocity variation dependent on depth alone can explain the existence of group velocities outside this range.

Low-velocity bottom.—Press and Ewing (1948b) have applied the normal-mode theory to the case where the bottom material has a lower speed than that of water. They show that such a configuration, with a low-velocity layer between two higher-velocity layers, should yield a dispersion curve having two branches. Along one of these, the high frequencies travel with the speed of sound in the low-velocity layer. Along the other branch, there is considerable attenuation of all frequencies except the highest, which arrive with the speed of sound in water. The pattern of high-frequency arrivals traveling at a lower speed than that of water has never been observed. High-frequency, nondispersive arrivals traveling with the water speed were obtained from all Bikini shots recorded with the detectors off the USS GILLIS when anchored near Cherry Island. These were shots no. 110 to 120 along the eastern part of the Enyu-Cherry profile. Since the same shots when received at Enyu give the usual dispersion pattern, there is probably a low-velocity bottom only under the western end of the trajectory.

Horizontal lithology change.—The normal-mode theory as applied to the propagation of explosion waves through the water has been developed on the assumption that the elastic properties of the bottom vary only with depth over the entire wave trajectory. Any horizontal

change in these properties would introduce new boundary conditions which could change the dispersion characteristics markedly and render the theory in its present form inapplicable. Possibly the theory can be extended to cover rigorously the introduction of a vertical discontinuity in speeds as a new boundary condition.

According to Emery's preliminary report (1948) on the bottom composition of Bikini lagoon, several zones of different bottom lithologies are concentric with the lagoon's boundary. The outermost ring covering the shallow terrace at the edge is coarse sand which grades lagoonward into fine sand. At 15 to 20 fathoms depth a disc of *Halimeda* debris begins, and this covers the entire central portion of the lagoon. No specific lithologic boundaries have yet been mapped, but the gradation from finer to coarser bottom sediments as one goes from the center to the periphery of the lagoon appears well established. Since all recording hydrophones were within 3000 feet of shore, it is expected that the receiving end of all wave trajectories would be in the relatively coarse material characteristic of the lagoon's edge. Where the explosion was in the deeper part of the lagoon, this end of the trajectory would have been over the fine sediments characteristic of the center. For these shots one would therefore expect some kind of horizontal lithologic change beneath the path of the water waves. Shots near the edge of the lagoon, on the other hand, would then have both the beginning and end of their trajectories over the same kind of bottom material, and one would expect the results to conform more closely to a theory assuming lateral homogeneity.

As a working hypothesis, let us assume that the discrepancies between the observed high-frequency dispersion and the theoretical predictions can be correlated with lithologic changes (down to about 100 feet below the bottom) between the two ends of the wave trajectory. It appears likely on the basis of available evidence that the bottom constitution is related to the depth of water and the distance from the lagoon's edge. It might be instructive to tabulate the high-frequency dispersion data with respect to these two variables. The data are separated according to shot location; the areas specified on the map (Pl. 1) by the Roman numerals are considered individually. In the following tabulation the agreement of the dispersion data with the refraction data is designated by one of three letters. *A* indicates close agreement between refraction and dispersion, (Fig. 6); *B* indicates poor agreement, but the difference is not so great that it could not conceivably be explained by experimental uncertainty (Fig. 7). *C* indicates that the observed dispersion is definitely outside the region which the normal-mode theory allows (Fig. 8).

Area	Recorded off Enyu (E) or Yuro (Y)	Average depth (fathoms)	Aver. distance from edge (naut. miles)	Agreement of dispersion data and refraction data
I	E	29	3.2	C
I	E	31	4.0	C
III	E	28	2.2	B
IV	Y	22	1.2	A
V	Y	29	2.0	C
VI	Y	30	4.5	B
VII	Y	27	2.7	A
VIII	E	22	1.1	A
IX	E	23	1.2	B
X	E	27	1.2	C

Thus better agreement is observed where the shots are in shallow water and near the edge. The principal exceptions are in areas VI and VII, which are in deep enough water and are close enough to the center of the lagoon that, on the basis of the results in other areas, one would not expect as good agreement as is obtained. In general the table supports the hypothesis that the better agreement is obtained where there is the least change in water depth and distance from the lagoon's edge between shot and detector.

The high-frequency dispersion data are applicable to depths of the order of 100 to 150 feet, while the Airy-phase data represent penetration to about 400 feet. The Airy dispersion exhibits no such variation with shot position but indicates bottom velocities that are uniform everywhere in the lagoon except along the southern edge. These velocities give good agreement with those that would be expected from the refraction data. Thus any lithologic change necessary to explain the discrepancy in the

high-frequency dispersion would not have to extend any farther than 100 feet below the lagoon bottom.

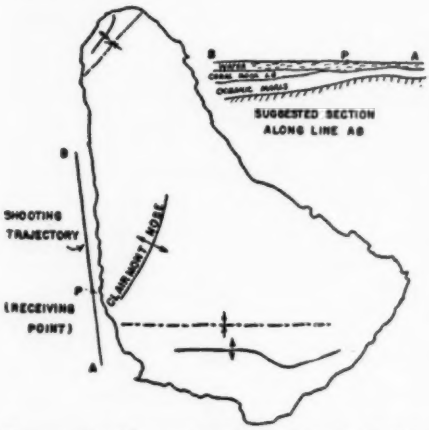

FIGURE 10.—MAP OF BARBADOS ISLAND, B.W.I.
Showing structural geology in region of shooting trajectory for Ewing and Worzel's under-water explosion measurements in 1944. Geologic section highly hypothetical.

To apply another test to our working hypothesis of horizontal change in bottom lithology, let us consider once more the bottom geology of the submarine shelf off Barbados where the same discrepancy is observed as with the deeper Bikini shots. Figure 10 shows the location of the shooting trajectory alongside the western edge of Barbados with respect to some of the structural elements in this area. All shots were received at point *P* on the southern end of the shooting trajectory; all those showing dispersion have wave paths 4 miles long or less. Point *P* appears to lie less than a mile north of the under-water extension of the Clairmont nose, a prominent anticiline. The hypothetical cross section for the region near the under-water detector illustrates how the Coral-rock formation might have been uplifted and eroded near the axis of the anticline so that the marly "Oceanic formation" might be the predominant bottom sediment. Farther north, down the limb of the anticline, the Coral-rock would reappear and extend from the water bottom to a depth of 100 feet or more. Thus there would be a lateral discontinuity between the bottom sediments at the shot and the detector ends. This hypothesis could be checked with deep-coring apparatus now available. Here are two widely separated areas for which there is indirect geologic evidence to explain anomalously low group velocities on the basis of horizontal lithologic change.

Those portions of the Bikini dispersion data to which the Pekeris normal-mode theory for liquid bottoms applies show that the average sound speed of the sediments in the uppermost 400 feet below the lagoon is 1.25 times that of sound in water. In the uppermost 20 feet the average ratio is about 1.05, while below 40 feet the average is 1.3. This distribution is similar to that obtained by Pekeris by analysis of Worzel and Ewing's dispersion data at Jacksonville (deep) where the water bottom consists of water-saturated sands increasingly compacted with depth.

Most of the shots fired in the deeper parts of the lagoon gave rise to Airy waves with periods that indicate bottom speeds comparable to those previously determined by refraction. The higher-frequency water waves from these shots generally had anomalously low group velocities from the standpoint of normal-mode theory. There appears to be a strong tendency for the areas where these shots were fired to have much finer-grained sediments below the shots than below the detectors. It is suggested that a lateral change in lithology is responsible for the anomalous group velocities observed from these shots. This hypothesis can be supported by indirect geologic evidence at Bikini as well as at Barbados, where the same discrepancy has been observed. Any such lithologic variation should extend downward only about 100 feet below the lagoon bottom since the Airy-phase data give no indication of such a variation below this depth.

If this hypothesis could be substantiated, it would indicate that the sedimentation in the lagoon may have followed its present pattern only for a limited time, namely that necessary to deposit the uppermost 100 feet or less of the sediments now below the lagoon. The gradation from edge to center of coarser to finer surface material indicates that the lagoon bottom may have been formed by wearing down of the reefs and islands around the periphery, probably by wave action. The difference between the high-frequency and Airy-phase dispersion patterns indicates the possibility of an interruption or

discontinuity in the process at 50–100 feet. Since the discontinuity does not appear to extend completely across the atoll, its sedimentary interpretation is not clear. Although the 1947 boring on Bikini island has thus far given no data on ages of formations less than 900 feet deep, it is possible that the inferred lithologic change 100 feet or less below the lagoon bottom is associated with the lowering of sea level to 300 feet below its present value during the Pleistocene glaciation, which Daly (1915) has called upon to explain the form of modern atolls.

If further investigations support the hypothesis of lateral lithologic change it might have important applications in the study of microseisms. Press and Ewing (1948a) have applied normal-mode theory to postulate that microseisms represent minimum or maximum group velocities of multiply reflected water waves that have traveled landward from storm centers over the deep oceans. The transmission of normal-mode water waves from the deep sea to land obviously involves a substantial horizontal change in bottom lithology at the continental shelf. The low speed (7000–12,000 ft/sec) in the first few thousand feet of bottom below the shelf surface has been well demonstrated in the series of offshore seismic measurements undertaken on the Atlantic shelf by Ewing *et al.* (1950 is the most recent publication) since 1937. This is in contrast to the much higher basement speed of 25,000 ft/sec recently measured by Ewing (1949) in some deep sea refraction work in the Atlantic.[5] When it becomes possible to test the normal-mode theory of microseisms more completely, it may be found that the horizontal lithology change introduces troublesome discrepancies of the type observed in the Bikini shooting. It is hoped that the present normal-mode theory can eventually be extended to cover this more complicated case.

[5] The difference between the 25,000 ft/sec basement speed reported by Ewing for the Atlantic Ocean and the 17,000 ft/sec speed reported in Part I for the speed of the deepest layer under Bikini might raise some question. It should be pointed out that the Bikini speed was measured along the top of a buried hill projecting 1–2 miles above the normal depth of the ocean bottom in this part of the Pacific. The material constituting the hill, presumably a magmatic intrusion through the original ocean floor, could well be of different composition from the normal igneous rocks constituting the Pacific Basin. Further, the precision of the Bikini speed measurement was such that it could be as much as 1500 ft/sec higher and still remain within the probable error.

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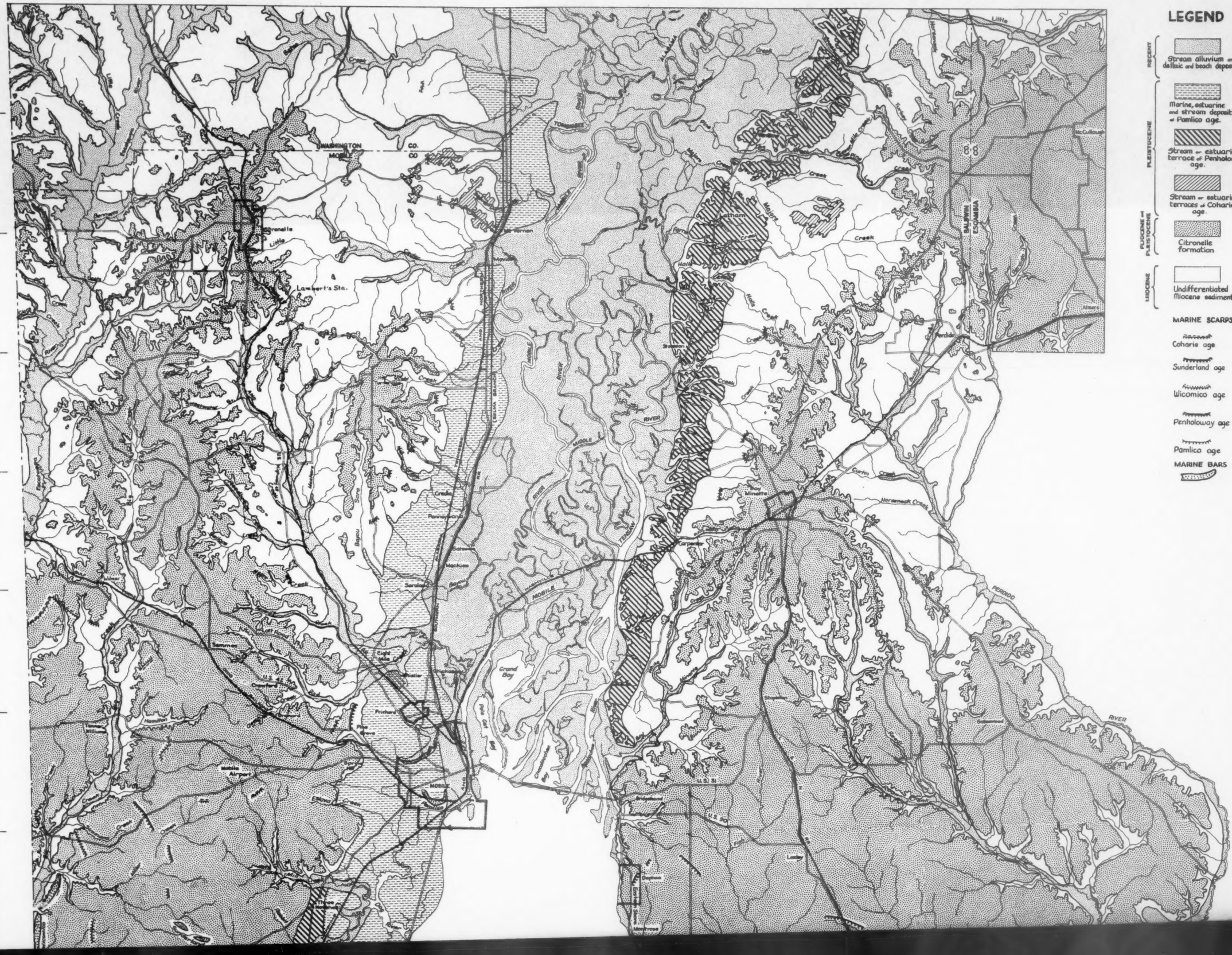
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GEOLOGIC MAP OF COASTAL ALABAMA

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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PLEISTOCENE HISTORY OF COASTAL ALABAMA

BY CHARLES W. CARLSTON

CONTENTS

	Page
Abstract	1119
Introduction	1119
Purpose of present investigation	1119
Acknowledgments	1119
The Citronelle formation	1119
General character and age	1119
Pleistocene land forms and geologic history	1121
Earlier studies	1121
Marine terraces	1122
Method of study	1122
Altitude distribution of bases of scarps	1122
Correlation of the marine terrace groups	1123
Character of the shore lines	1124
Age of the terraces	1125
Citronelle-Coharie interval	1126
Fluvial and estuarine terraces	1126
Terraces of Coharie age	1126
Terraces of Penholoway age	1126
Terraces of Pamlico age	1127
Submerged valley of Mobile Bay	1128
Summary and conclusions	1128
References cited	1128

ILLUSTRATIONS

Figure	Page
1. Projected profiles of the surface of the Citronelle formation, southwest Alabama	1125

Plate	Facing page
1. Geologic map of coastal Alabama	1119
2. Distribution of marine scarps on the east Gulf coast shown on topographic maps	1122
3. Graph of elevation of marine terraces at base of scarp, from Lucedale, Mississippi quadrangle to Tallahassee, Florida quadrangle	1126

ABSTRACT

Following its deposition, the late Pliocene or early Pleistocene Citronelle formation was entrenched by consequent streams and then tilted toward the Gulf. Submergence in waters 190 to 210 feet above present sea level then resulted in a compound shore line and marine erosion of the Coharie terrace. Four other marine terraces occur below the Coharie level: the Sunderland at 150 to 160 feet, the Wicomico at 90 to 110 feet, the Penholoway at 60–70 feet, and the Pamlico at 20 to 30 feet. The Coharie, Penholoway, and Pamlico marine terraces are associated with fluvial or estuarine terraces in the Mobile River Valley. After Pamlico submergence, a Mobile River floodplain and delta was formed at 10 to 11 feet below present sea level.

INTRODUCTION

Purpose of Present Investigation

During the summer of 1945, while employed by the Ground Water Division, U. S. Geological Survey, the writer studied the post-Miocene sediments and landforms of coastal Alabama (Pl. 1). He mapped and studied the Pliocene or Pleistocene Citronelle formation, but the present paper principally discusses the post-Citronelle Pleistocene terrace deposits and landforms.

Acknowledgments

The writer is indebted to Watson Monroe and John T. Hack of the U. S. Geological Survey, and Richard Foster Flint of Yale University for helpful suggestions. A special note of thanks is due Watson Monroe who has critically read the finished manuscript and whose suggestions have been most helpful in all stages of its preparation. Responsibility for interpretations and conclusions reached in the paper is, of course, solely the author's.

THE CITRONELLE FORMATION

General Character and Age

The Citronelle formation and underlying beds constitute the bedrock exposed to post-

Citronelle stream and marine erosion and Pleistocene deposition. The Citronelle overlies beds as young as Eocene, but in the coastal area it overlies Miocene and possibly some Pliocene deposits. In the most recent work, F. S. MacNeil (1946), divided the Miocene of Alabama into the Paynes Hammock sand, fossiliferous, greenish to gray sand and clay, and the Catahoula sandstone which consists of variegated clay, sand, and sandstone, though predominantly gray, and most commonly a clay. The Paynes Hammock sand is regarded as the marine equivalent of the Catahoula sandstone. MacNeil does not recognize in Alabama the Miocene Hattiesburg clay and Pascagoula clay which overlie the Catahoula sandstone in Mississippi.

The Citronelle formation (Pl. 1) ranges from 40 to 130 feet thick in coastal Alabama, because of pre-Citronelle relief and a general thickening toward the Gulf. Relief increases rapidly back from the coast, amounting to about 300 feet at Citronelle, Alabama. Consequently, except for a narrow coastal strip, the Citronelle has been cut through to the underlying Miocene beds. The predominantly sandy Citronelle above the predominantly clayey Miocene beds results in steep slopes wherever dissection has laid bare the Miocene. Another result of this contrast is the development west of the Mobile River in Alabama of a strong Citronelle cuesta which faces north over a local peneplane on the Miocene clays. The distribution of the Citronelle formation in coastal Alabama is shown on the geologic map (Pl. 1).

The Citronelle dips south in Alabama at 6 to 8 feet per mile while its surface slopes 5½ feet per mile. In general, it consists of a thin basal gravel overlain by a thick sand section and capped with variegated clays. The sands are locally cross bedded and the clays are generally highly lenticular. The clay lenses may occur very close to the base, and gravel sheets and channel fills are scattered through the formation, locally capping it. It is virtually impossible to differentiate Pleistocene sand and gravel of the marine terraces from the Citronelle sand and gravel.

The gravel of the Citronelle—predominantly quartz and quartzite pebbles—was transported by the Alabama River from the Precambrian and Paleozoic areas of northern Alabama. West of the Mobile River, in the area around Citronelle, Alabama, Citronelle gravel is largely angular, brown to tan chert pebbles. In Mississippi, pebbles are almost entirely of chert. Since there is no adequate source of chert in the watersheds of the streams now crossing the Alabama and Mississippi coastal plain, it is suggested that the Mississippi River contributed some sediments to the formation in Alabama.

No invertebrate or vertebrate fossils have been found in the Citronelle formation. Fossil leaves dated as late Pliocene by Berry (1916, p. 193–208) were supposedly collected from the Citronelle formation, but one of the two collecting localities was actually from clays underlying the true Citronelle (Roy, 1939, p. 1553–1559). Beds which appear to be of the same age as the Pliocene Charlton formation of Georgia were encountered at 59 feet below sea level in a well at Mobile, Alabama, and the base of the Citronelle is 80 feet above sea level at Mobile. However, Mobile lies in a broad and deep valley cut into the Citronelle; therefore, the Pliocene beds could have been deposited in an earlier estuary cut back into the Citronelle formation and thus may postdate the Citronelle. G. F. Brown (Brown, *et al.*, 1944, p. 54) found invertebrate fossils between the Citronelle and the Miocene, in beds which he named the Graham Ferry formation. The fossils were identified by Julia Gardner of the U. S. Geological Survey as forms which range from the Pliocene through the Pleistocene. Thus the Graham Ferry formation could be as old as early Pliocene or as young as Pleistocene. The evidence of the Graham Ferry formation taken with the presence of Pliocene beds in the well at Mobile suggests that the Citronelle is no older than late Pliocene.

The writer has found evidence of five marine terraces at elevations up to 210 feet above present sea level which cut the Citronelle formation in Alabama and adjacent Mississippi. Although there is no evidence that the age of the terraces extends back to the early Pleistocene, they suggest that if the Citronelle formation is Pleistocene it must have been deposited comparatively early in the Pleistocene. If the heavy aggradation by the lower Mississippi River and

Mobile River which formed the Citronelle formation was due to uplift in the interior of the continent, the Citronelle could be either late Pliocene or early Pleistocene. If, however, the cause was climatic, the known change of climate which brought on Pleistocene glaciation in the higher latitudes could account for its deposition. The advent of the Pleistocene may have brought a greatly increased precipitation to the southern states.

If the Citronelle formation is late Pliocene in age, it obviously does not enter into a discussion of the Pleistocene history of coastal Alabama. If, however, it is earliest Pleistocene in age, the Pleistocene history of coastal Alabama properly begins with the aggradation of the Mississippi and Mobile Rivers building great alluvial fans at the continental margins forming the Citronelle formation.

Pleistocene Land Forms and Geologic History

Earlier Studies

Matson (1916, p. 180–190, pl. 32) recognized four "plains" and three Pleistocene terraces forming distinct surfaces on the Citronelle formation of coastal Alabama. The "plains" (1916, p. 180) are described as divisible into stream terraces and interstream plains; some parts of the interstream plains are described as fluviatile in origin and some parts as marine. The four plains in western Alabama are, from oldest to youngest, the Brookhaven plain, between elevations of 420 and 350 feet, the highest surface on the Citronelle; the Sardis plain, from 350 to 310 feet; the Canton plain, from 310 to 250 feet; and the Loxley plain, from 250 to 200 feet. The highest elevation for each plain is stated to be at the inner margin of the plain and the lowest elevation at its outer margin. Inasmuch as the outer margin on each terrace is stated to be at the same elevation as the inner margin on the next lower terrace, the range in elevations given by Matson suggests a smooth slope rather than steplike terraces. The problem of finding these terraces is further complicated by Matson's apparent errors in elevation. For example, in western Alabama west of the Mobile River, the Citronelle formation in the vicinity of Citronelle reaches a maximum elevation of only 349 feet.

The terraces of Pleistocene age recognized in western Alabama by Matson were, from oldest to youngest, the St. Elmo, the Hammond, and the Pensacola terraces. According to Matson's Plate 23 (1916), the St. Elmo terrace is at about 175 feet, the Hammond at about 40 feet, and the Pensacola terrace at about 25 feet. These terraces (1916, p. 189) are described as extensive and as having seaward-facing scarps.

G. F. Brown (Brown, *et al.*, 1944, p. 19) states that east of Pascagoula County in George County, Mississippi, there are three benches on this upland, at altitudes of 200–230 feet, 260–280 feet, and 300–310 feet. The 300–310 foot bench is stated to reach an altitude of about 340 feet near Citronelle, Mobile County, Alabama. He notes the presence of two stream terraces and one marine terrace, the Pamlico terrace. At 42 feet above sea level in a beach gravel, he found flattened pebbles with mollusk borings, but could not positively correlate the gravels with Cooke's Pamlico terrace, or Talbot terrace or some older terrace (Brown, *et al.* 1944, p. 22, 23).

Cooke (1945, p. 12–13, 273–312) recognizes seven Pleistocene marine terraces in western Florida: the Brandywine terrace at 270 feet, the Coharie terrace at 215 feet, the Sunderland terrace at 170 feet, the Wicomico terrace at 100 feet, the Penholoway terrace at 70 feet, the Talbot terrace at 42 feet, and the Pamlico terrace at 25 feet. The probable limit of error in these elevations is given as 5 feet (Cooke, 1945, p. 12). The seven shore lines are shown on a series of maps, but since the scale is about one to eight million, they are of little value in locating the shore lines in the field or on topographic maps. They also show the shore lines as continuous lines, not differentiating mappable scarps from inferred locations of the shore lines. Locations of scarps are described by counties in the text discussion of each shore line.

Vernon (1942, p. 15–28) describes five "coastwise" surfaces and four fluvial terraces in Holmes and Washington counties, Florida. The coastwise surfaces are named only by elevation but are correlated with Cooke's shore lines. The

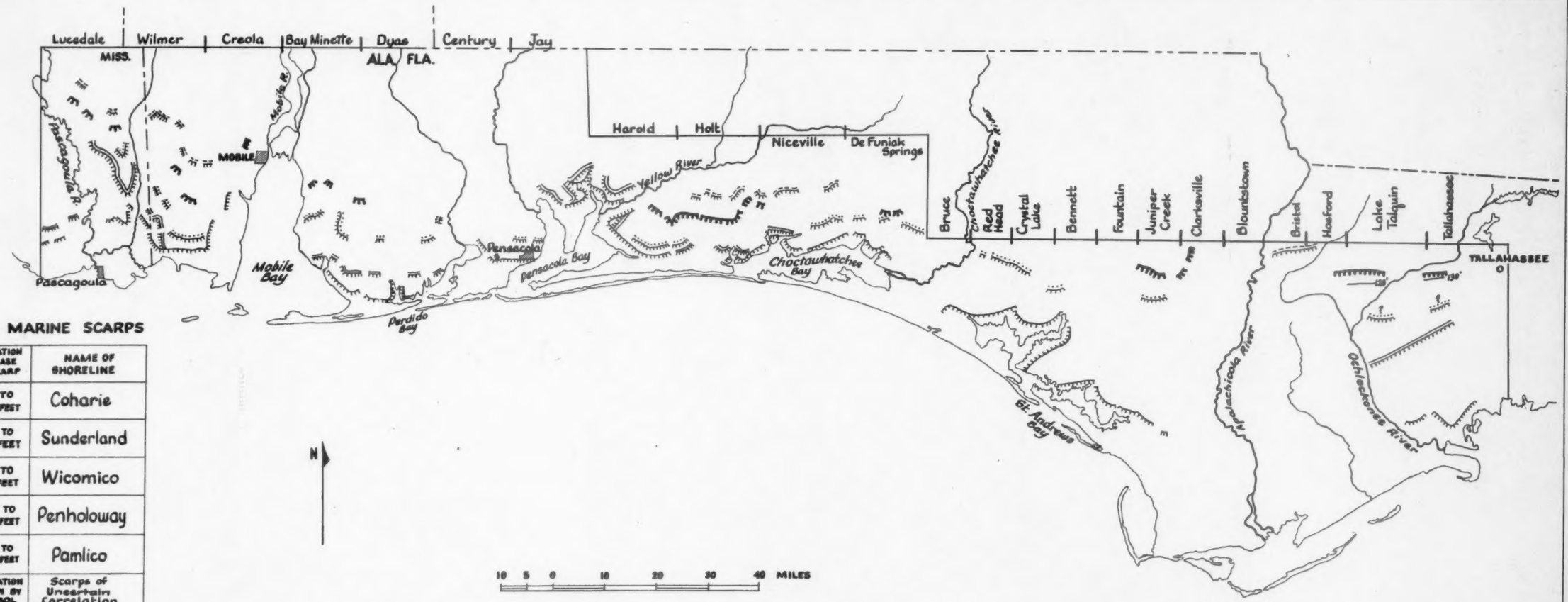
surfaces and their correlations (Vernon, 1942, table 2) follow: 250–320 feet, Brandywine and Citronelle; 170–220 feet, Coharie; 115–150 feet, Sunderland; 60–105 feet, Wicomico; 5–30 feet, Pamlico. Each surface, except the highest, is bounded by a marine scarp, whose elevation, at base of scarp, is the higher of the two elevations given for each surface. Thus there are four marine scarps or shore lines described, no scarps being found above 220 feet in elevation. No evidence was found of Cooke's Brandywine (270 feet), Penholoway (70 feet), and Talbot (42 feet) shore lines. The fluvial terraces are also named only by elevations, but are correlated with alluvial terraces described by Fisk (1939) in Louisiana. Their elevations and correlations follow: 250–320 feet, Williana; 145–165 feet, Bentley; 60–100 feet, Montgomery; 30–50 feet, Prairie; 10–20 feet, not recognized by Fisk. The fluvial terraces are also correlated with the coastwise surfaces. An excellent map (Vernon, 1942, fig. 7) shows the locations of the coastwise terrace scarps in southern Washington County.

Marine Terraces

Method of Study.—In the course of field work on the Alabama coast a number of scarps were noted in the field and identified on topographic maps or, more often, noted on topographic maps and located in the field. Except for the excellent wave-cut cliff of the Pamlico terrace at about 25 feet above tide, all the stronger scarps that stood out clearly in the field or on topographic maps were parallel with the south-flowing streams that dissect the area, or were more nearly parallel with the streams than with the coast line. For that reason it was at first believed that the scarps were all stream cut, delimiting stream terraces of varying altitudes. Since the field work was in part financed by the State Geological Survey of Alabama, it was not possible to carry out an extensive search for marine terraces outside the boundaries of the state. Field work inside the state boundaries failed to disclose convincing evidence of marine terraces above an altitude of about 25 feet, the level of the Pamlico shore line. It was noted, however, that Flint (1940; 1947) had been unable to find any evidence of five of Cooke's seven shore lines along the Atlantic coast, yet Flint and Cooke were in agreement that the 100-foot Wicomico shore line and 25-foot Pamlico shore line were valid shore lines. Flint pointed out in a letter to the writer that he had seen evidence of terraces higher than 25 feet while flying over west Florida. The writer, therefore, began a study of recently published U. S. Geological Survey topographic maps which cover an area extending eastward in a continuous strip along the Florida coast from the Alabama line to a short distance beyond Tallahassee, Florida. While at first searching for evidence of the 100-foot Wicomico shore line, the writer soon became aware that there existed a number of other escarpments generally parallel to the coastline and apparently of marine origin. These were located on a base map of scale 1:1,000,000 (Pl. 2). The elevations, to the nearest 10-foot contour interval, of the base of the scarps were plotted on a graph (Pl. 3). Two projected profiles were constructed parallel to the dip of the Citronelle and normal to the coast line on the surface of the Citronelle in order to help locate the terrace scarps in Alabama (Fig. 1).

Altitude distribution of bases of scarps.—Examination of the altitudes of the bases of the many scarp remnants (Pl. 3) shows that there are scarps at 19 of the 22 10-foot intervals between sea level and 220 feet, the highest elevations at which the scarps were found (Pl. 3). Only at 10, 80, and 140 feet were there no scarps. Such a distribution in elevation of marine terrace scarps might indicate little more than a series of scarps cut during the slow retreat of the sea from 220 feet above present sea level to its present level.

Examination of the total number of miles of scarp at each 10-foot interval showed, however, that there is a grouping of levels at 20 to 30 feet, 60 to 70 feet, 90 to 110 feet, 150 to 160 feet, and 190 to 210 feet. The groupings of terrace levels thus have a range of 10 to 20 feet; the spaces between the groups have a range of 20 to 40 feet. Thus the spaces between terrace groups at minimum are equal to the range of terrace groups and at maximum are twice the range of the terrace groups. This ratio between ranges of elevations of the terrace groups and ranges of elevations between the groups appears to favor the hypothesis that there are five groups of terrace remnants



LEGEND FOR MARINE SCARPS

SYMBOL	ELEVATION OF BASE OF SCARP	NAME OF SHORELINE
	190 TO 210 FEET	Coharie
	130 TO 160 FEET	Sunderland
	90 TO 110 FEET	Wicomico
	60 TO 70 FEET	Penholoway
	20 TO 30 FEET	Pamlico
120'	ELEVATION SHOWN BY SYMBOL	Scarps of Uncertain Correlation

RECTANGULAR LINES INCLOSE AREA TOPOGRAPHICALLY MAPPED BY THE U.S. GEOLOGICAL SURVEY

DISTRIBUTION OF MARINE SCARPS ON THE EAST GULF COAST SHOWN ON TOPOGRAPHIC MAPS

along the East Gulf Coast which can be correlated through that distance. The ratio is not sufficiently high, however, to constitute evidence that the hypothesis is proved.

which has significance with respect to former sea level is the planed rock surface of the inner edge of the wave-cut platform, at the base of the wave-cut cliff beneath its cover of alluvial

TABLE 1. COMPARISON OF TERRACES FOUND BY WRITER AND MARINE TERRACES DESCRIBED BY COOKE (1945) AND VERNON (1942)

Present Investigation	Cooke, 1945	Vernon, 1942
Not recognized	270-feet, Brandywine terrace	Not recognized
Scarps at 190–210 feet	215 feet, Coharie terrace	220-foot terrace
Scarps at 150–160 feet	170 feet, Sunderland terrace	165-foot terrace
Scarps at 90–110 feet	100 feet, Wicomico terrace	100-foot terrace
Scarps at 60–70 feet	70 feet, Penholoway terrace	Not recognized
Not recognized	42 feet, Talbot terrace	Not recognized
Scarps at 20–30 feet	25 feet, Pamlico terrace	30-foot terrace

It is improbable that the original elevations of the base of each scarp were precisely equal. Variations in the height of storm terraces built up against the wave-cut cliff, variations in tidal range as affected by coastal configurations, and the unknown effects of hurricanes along the Gulf Coast, which have been observed to raise the height of wave attack from 8 to 15 feet above normal Gulf level on shores near the center of the storms (Cline, 1920, p. 127), are all factors which might or actually do vary the original elevation of the base of wave-cut cliffs. Far more important is the effect of varying thicknesses of small alluvial fans, slope wash, and colluvium in changing the elevation of the base of the original scarp since its date of cutting. For example, at a number of places along the 20- to 30-foot terrace, the base of the scarp follows the 30-foot contour in stretches facing the sea, and then changes to the 20-foot contour where the scarp bends northward into an estuarine embayment. The slope of the upland above the 20–30 foot terrace is southward and therefore slope wash and stream transportation of sediment eroded from the land above the wave-cut cliffs will be more greatly directed southward over the cliffs and to the base of scarps running in an east-west direction, or parallel to the coast, than over the edge of scarps running parallel to the regional slope. Thus even the most recent marine terrace has been built up at its scarpward margin by slope wash and stream deposition.

As Douglas Johnson (1944, p. 801) has pointed out, the only part of a marine terrace

fans, colluvium, and beach debris. The similarity between the wave-reworked Citronelle of the old terraces and the bed-rock Citronelle in Alabama is so great it was found impossible to determine these altitudes on the marine terraces in Alabama. Where the bed rock of the wave-cut platforms of the marine terraces is sufficiently different lithologically from the old beach deposits and colluvium, conditions which probably exist farther east from the Alabama-Florida line, such elevations must be determined before the terrace groups can be evaluated as precisely-correlative terraces.

Correlation of the marine-terrace groups.—Neither the writer nor Vernon was able to find the 270-foot Brandywine shore line recognized by Cooke (Table 1). The 190- to 210-foot group of terraces corresponds fairly well with Cooke's Coharie terrace, and the 150- to 160-foot group shows about the same degree of correspondence with Cooke's Sunderland terrace. The scarps at 90 to 110 feet agree well with Cooke's 100-foot Wicomico terrace, as does the 60–70 foot group with Cooke's Penholoway terrace and the 20- to 30-foot group with Cooke's Pamlico terrace. Cooke's 42-foot Talbot terrace does not appear to be well represented since there are only about 3 miles of terrace scarp at 40 feet and one of the 3 miles is a higher portion of the 20- to 30-foot terrace. The correspondence with Vernon's marine terraces in Holmes and Washington County, Florida, is of high order since neither the writer nor Vernon found evidence of the 60–70 foot group of terraces in Washington County, Florida.

Unfortunately all of Cooke's shore lines except the Wicomico shore line are mapped on far too small a scale, about one to eight million (Cooke, 1945), to permit comparison with the terrace scarps found by the writer along the west Florida coast. The 100-foot Wicomico shore line is given in greater detail on the Geological Map of Florida by Cooke (1945, Pl. 1). Of the approximately 51 miles of scarps at 90 to 110 feet which were mapped by the writer in west Florida, only 6 miles of scarp coincide with Cooke's map of the Wicomico shore line.

In strict accuracy the terraces should therefore be referred to in terms of their elevations, or given local names. The first method, however, seems unnecessarily cumbersome, and the second would only add more names to the already overnumerous list of terrace names. The writer therefore adopts Cooke's nomenclature, with the provision that correlation is not considered to be proved, and that the names only represent a hypothesis that the correlation exists. The equivalence of terrace groups with Cooke's shore lines rests as shown in Table 1.

Character of the shore lines.—An effective method of analyzing the seaward slope of the surface of the Citronelle formation in Alabama was suggested to the writer by John T. Hack of the U. S. Geological Survey (personal communication). The method essentially consists of constructing a projected profile over a strip of land up to several miles wide. A simple profile along a straight line is constructed parallel with the regional dip of the formation or regional slope of the surface. A line is then drawn parallel with the first profile about a fourth to half an inch from it. The second line of profile is then projected on to the original profile so that no elevations of the second line which are below the first line of profile are shown. All elevations of the second line which are higher than those along the first line replace the elevations of the first line. This is done along further equally-spaced profile lines ending with a composite profile of a strip of country a mile to several miles wide. The writer used this method in constructing profiles with half mile spacing and 3-miles width. The center line for the first profile is the 87° 45′ line of latitude which is located in Baldwin County, east of Mobile Bay. The center line for the second profile is the 88° 15′ line which is in Mobile County, west of Mobile Bay. Elevations at half-mile intervals are plotted at .1-inch intervals (Fig. 1).

On the Baldwin County profile, the Pamlico scarp is well shown at 20 to 30 feet. The Penholoway terrace shows up well at about 55 feet, the Wicomico at about 105 feet. The Sunderland scarp is fainter, but appears to be present at 145 feet. Dissection has removed any possible evidence of the Coharie scarp at 200 to 220 feet.

The Mobile County profile shows a very strong wave-cut cliff with a base at about 20 feet, or at the Pamlico terrace level. The Penholoway scarp is missing, but the Wicomico terrace appears at about 95 feet, the Sunderland at about 150 feet, and the Coharie is well shown at 200 feet.

The striking difference in prominence of the wave-cut scarp of the Pamlico terrace east and west of Mobile Bay is due to the protection afforded the Pamlico shore line east of Mobile Bay by a marine bar similar to the modern bar which skirts the Baldwin County shore. The Pamlico bar is about a mile and a half inland from the modern bar and extends up to 30 feet above sea level, although occasional dunes which top it reach up to 40 feet elevation. Since the scarps above 30 feet on both profiles are comparable in prominence to the bar-protected Pamlico scarp of Baldwin County, the suggestion is strong that these higher scarps, if valid marine scarps, were also so protected from active marine erosion at the time of their formation.

The trend of the scarps (Pl. 2) clearly indicates the presence of drowned valleys during cutting of all five of the shore lines. The shore line during erosion of the terraces appears to have been of the same type that it is now. At present, the drowned valleys of Mobile Bay, Perdido Bay, Pensacola Bay, Choctawhatchee Bay, and St. Andrews Bay coincide with the mouths of the larger streams which enter the Gulf. Between these streams and bays, relatively less dissected stretches of the coastal slope are faced with a fine off-shore bar. The present coast line, therefore, shows in marked development characteristics of both a shore line of submergence and a shore line of emergence, thus making it a compound shore line.

At present, the off-shore bar protects the land directly behind it from wave erosion. Within the bays, however, wave erosion continues, though greatly lessened by the short fetch of open water. On the smaller bays of this stretch of coast, wave erosion is probably relatively insignificant; but on such a body of water as Mobile Bay, wave erosion may be observed actively wearing back the shores of the bay.

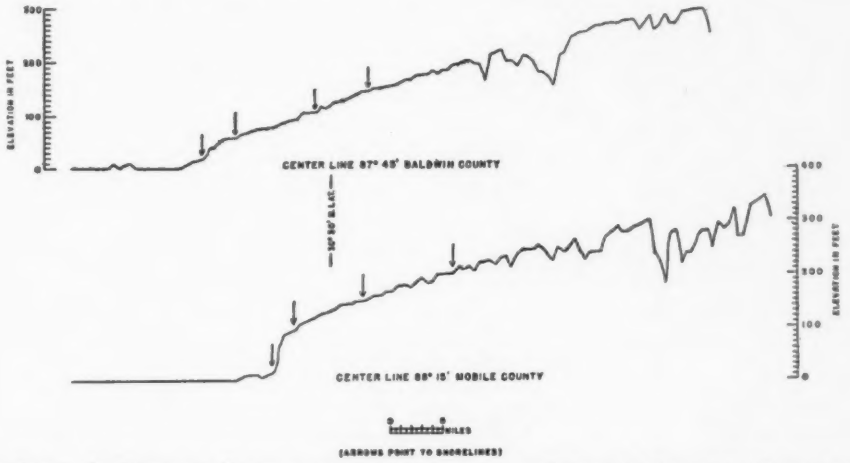


FIGURE 1.—PROJECTED PROFILES OF THE SURFACE OF THE CITRONELLE FORMATION, SOUTHWEST ALABAMA

As previously described, the marine scarps facing the Gulf in Alabama are very faint, suggesting that they were protected from more active erosion by off-shore bars. The scarp segments which face the present stream valleys, or the old estuaries, are much higher and well-defined. This is entirely expectable in the light of present conditions in Mobile Bay.

The faintness of the Gulf-facing marine scarps in Alabama contrasts strongly with the much higher and better defined marine scarps of the five shore lines east of Pensacola Bay. In this stretch of the Gulf Coast, the coastal slope was either too steep for development of an off-shore bar, or more likely, the shore lines had been eroded back to a stage of submaturity, at which stage the off-shore bar had been driven back against the headlands of the inner shore.

Age of the terraces.—Cooke (1945, p. 17, 248) correlates the Brandywine shore line with the Aftonian interglacial stage, the Coharie and Sunderland shore lines with the Yarmouth stage, the Wicomico, Penholoway and Talbot shore lines with the Sangamon stage, and the Pamlico with the Wisconsin. Flint (1947, p. 270–271) tentatively correlates the Surry scarp (Wicomico) with the Yarmouth stage and the Suffolk scarp (Pamlico) with the Sangamon.

The highest terrace scarp, the Coharie, appears somewhat more fragmentary and less continuous than the Sunderland terrace; while the Sunderland and Wicomico terraces show a similar degree of continuity and erosional modification. The difference in degree of preservation of the Coharie, Sunderland, and Wicomico scarps is not, however, particularly evident. The Penholoway and Pamlico scarps appear to have a like degree of freshness and the marine bars on the two terraces show a very similar degree of preservation. The only observable difference between the two scarps is that the Pamlico scarp is considerably more continuous than the Penholoway scarp. If the Penholoway and Pamlico scarps were cut during the same Pleistocene stage, as appears distinctly possible, the difference in their continuity could be due to the difference in the length of time the sea stood at the two levels. This may also explain the difference in degree of continuity between the Coharie scarp and the two next lower scarps.

Citronelle-Coharie interval.—Since even the highest shore line, the 200- to 220-foot Coharie shore, outlines a series of submerged valleys, it must be concluded that the Citronelle formation had been subjected to subaerial erosion previous to the time when the first marine terrace was cut in its coastal slope.

It is difficult to distinguish the effects of pre-Coharie stream erosion from post-Coharie erosion except at two localities in southern Alabama. The first of these, 2 miles west of Wilmer, Alabama, (Pl. 1) is shown on the Wilmer, Alabama, quadrangle. Here the Coharie scarp, at 220 feet, is cut back in a sharp re-entrant. At the apex of the re-entrant, an abandoned stream valley is cut through the Citronelle upland. The stream which cut this valley may have been the upper Escatawpa River, which is in direct alignment with the valley to the north, or it may have been a short stream which was captured through intercision by the Escatawpa. The valley is about half a mile wide and about 40 feet deep below two hills which flank it.

The second locality is east of Mobile Bay on the Fairhope quadrangle, 2 miles northeast of Daphne (Pl. 1). Here a broad, shallow valley has resulted in well-defined re-entrants in the Wicomico and Sunderland scarps. The valley appears to have been formed by a fairly small stream, is up to 6 miles wide, although it is only about 10 to 20 feet deep. Although the valley is too low in elevation to be nipped by the Coharie scarp it too may have been formed in the post-Citronelle, pre-Coharie interval.

In the following pages it will be shown that stream or estuarine terraces of Coharie age extend for miles up the Mobile River valley. The present course of the Mobile River was, therefore, established during the Citronelle-Coharie interval.

A third event of the Citronelle-Coharie interval was tilting of the Citronelle formation to its present slope of about 5.5 feet per mile toward the Gulf. The tilting may have involved positive upwarping of the land or a differential warping that raised the landward part of the Citronelle and lowered the seaward margins. The uplift is referred to the Citronelle-Coharie interval under the hypothesis that the Coharie terrace is an unwarped and undeformed datum plane.

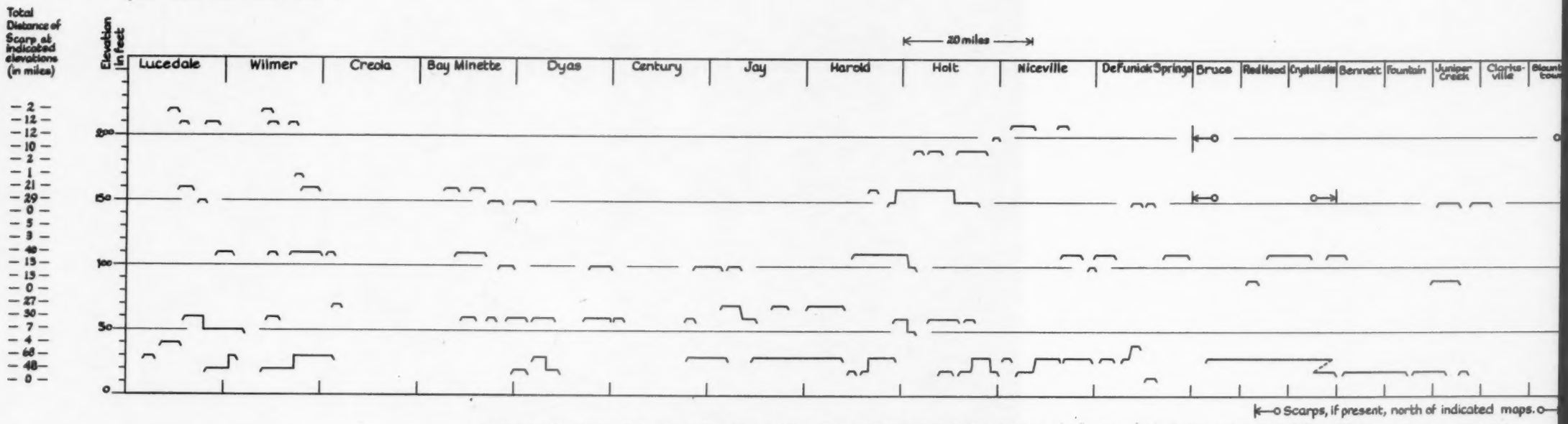
The slight depth of incision by streams in the two areas described above suggests that the erosion occurred before tilting of the Citronelle. There is not sufficient data at hand, however, to determine whether that erosion reached a stage of youth, maturity, or post-maturity. Neither is it possible to determine positively the depth of incision by streams following the tilting, but before Coharie marine erosion.

Fluvial and Estuarine Terraces

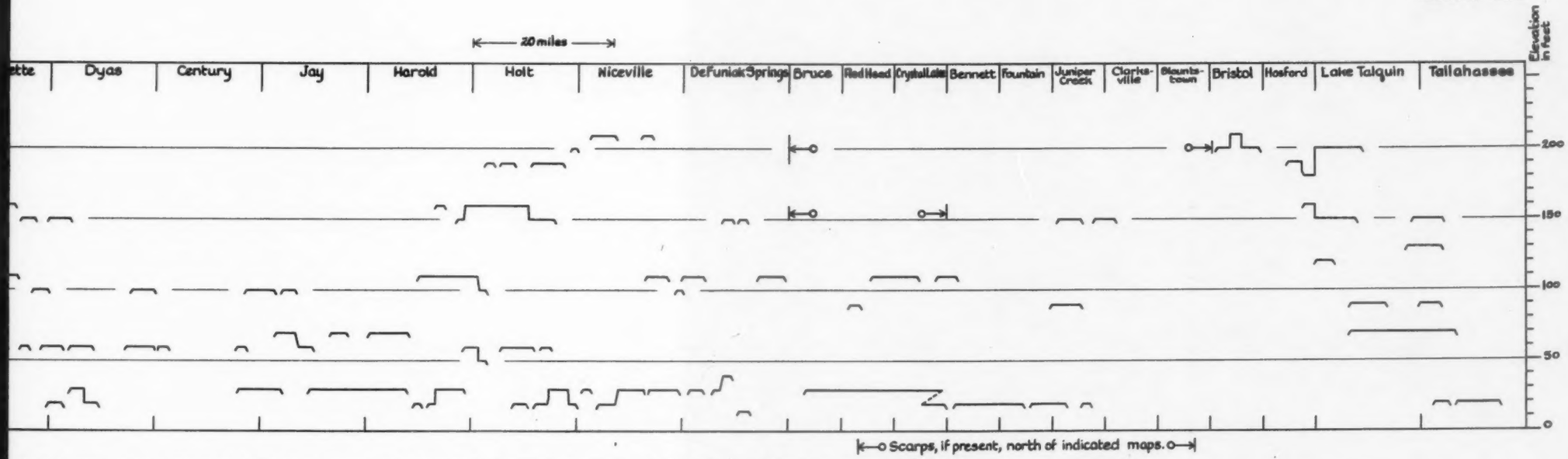
Terraces of Coharie age.—A series of scarps at 190 to 210 feet in elevation are mapped (Pl. 1) along the upper reaches of the Mobile River Valley sides. The scarps face toward the valley and were cut either by wave action in a long estuary extending up the Mobile Valley or by the Mobile River when it was graded to the level of the Coharie sea. Several terrace remnants are preserved at this elevation along the west side of the valley for 4 miles north from Mt. Vernon, Alabama. The gravel underlying the terraces is about 13 feet thick, which suggests that the terraces are of fluvial origin. A terrace remnant, less than half a square mile in area, is preserved at an elevation of 22? feet in the Tensaw quadrangle about 7 miles east of Latham. This terrace is about 10 to 15 feet higher than the terraces on the west side of the valley, yet its elevation is sufficiently close to the terraces near Mt. Vernon to be included as a terrace of probable Coharie age.

Terraces of Penholoway age.—The next lower surface below the Coharie terraces is a stream terrace confined to the east side of the Mobile River Valley which extends from Bay Minette Creek, at the northeastern end of Mobile Bay, some 40 miles north to Little River (Pl. 1). At Tensaw, Alabama, the surface of the terrace is at 140 feet. At its southern end, 31 miles downstream from Tensaw, its surface is about 65 feet, a slope of about 2.4 feet per mile. The lower end of the terrace is nearly horizontal, which suggests that its base-level was somewhere near 65 feet in elevation, or the level of the Penholoway marine scarps.

The town of Theodore, southwest of Mobile, lies on a small terrace remnant (Pl. 1). At Theodore the terrace is at an elevation of 55 to 65 feet; 2 miles north it is a little over 70 feet in elevation. These altitudes suggest that it too



GRAPH OF ELEVATION OF MARINE TERRACES AT BASE OF SCARP, FROM LUCEDALE, MISSISSIPPI QUADRANGLE TO TALLAHASSEE, FLORIDA QUADRANGLE



GRAPH OF ELEVATION OF MARINE TERRACES AT BASE OF SCARP, FROM LUCEDALE, MISSISSIPPI QUADRANGLE TO TALLAHASSEE, FLORIDA QUADRANGLE

is of Penholoway age. A long marine bar extends for about 5 miles south from the terrace. Where it joins the terrace a mile east of Theodore it is 55 feet in elevation. At its distal end, it is about 40 feet in elevation. How much of the bar was above water at the time of its formation cannot be exactly determined, but an elevation of about 50 feet appears to be within 5 feet of the actual sea-level of that time. The terrace at Theodore was probably at least partly of estuarine origin although its apparent lack of horizontality in a north-south direction weighs against this hypothesis. The relations of the bar to the terrace at Theodore suggest either that the Penholoway sea level was nearer 50 feet than the 60- to 70-foot elevation indicated by the marine scarps, or that there was an intermediate stand of the sea between 60 to 70 feet and 20 to 30 feet.

Terraces of Pamlico age.—A terrace of the Mobile River of Pamlico age extends along the west side of the valley from the northern end of the mapped area southward beyond the city of Mobile (Pl. 1). At Mt. Vernon, Alabama, it is about 50 feet above tide and its coastward slope declines to about 28 feet at Mobile. A recurved spit occurs on the terrace in the city of Mobile. The top of the spit is a little over 30 feet in elevation. Sixteen miles farther south, or 3 miles north of the present shore line, a second spit extends eastward 4 miles from an east-west-trending wave-cut cliff facing Mississippi Sound. The base of the cliff is about 25 feet above tide. The crest of the spit extending out from the cliff is at 32 feet. The river terrace thus becomes an estuarine terrace at its lower end, which in turn grades into a marine terrace about 25 feet above sea level.

Smith, Johnson and Langdon, (1894, p. 49) give a section of materials under the Pamlico terrace surface which were exposed by a storm in 1893. The section was at the Toll House "a mile or two" below Mobile. They report 6 to 7 feet of yellowish sand and some clay below the surface. Below the sand and clay is 6 to 12 inches of lignitic matter underlain by 6 feet (to tide level) of blue clay containing many rotten, partly bituminized wood fragments. Similar blue clays with decaying wood are penetrated by virtually every well sunk in downtown Mobile.

Beneath the blue clay occurs a thick section of Pleistocene fill whose exact depth is difficult to calculate. Although many water wells and test holes for water development have been drilled throughout the city and its suburbs, the well logs are very generalized and are by no means accurate. One of the better logs shows, for example, clay to 8 feet, "soapstone" from 8 to 21 feet, and sand from 21 to 91 feet. Below 91 feet is clay, shale, and "soapstone," with some interbedded sands. In eight other well logs, the bottom of the thick sand zone is given at depths of 64, 80, 85, 89, 96, 97, 103, and 106 feet below the terrace surface. Only a part of these differences is accounted for by the difference in elevation of the well heads, which range from 27 to 39 feet above sea level. The best available log was made by P. E. LaMoreaux of the U. S. Geological Survey on a water well drilled at the G. M. & O. Building in Mobile. Correlations of the log are by the present writer.

Log of G. M. & O. Water Well Drilled in 1943 by Carlos Well & Pump Supply Co.
(Measurements taken from rotary table which was 10 feet above sea-level)

Pleistocene alluvium	
0–14	Fill.
14–17	Sand, wood fibers, marsh remains.
17–30	Sand, light yellow, medium-grained, micaceous, and yellow clay particles.
30–48	Sand, white, coarse, slightly micaceous, some gravel; contains scattered red clay particles.
48–54	Sand, gray-white, coarse to gravelly, scattered red clay particles.
54–59	Same. Increase in fine gravel.
Pliocene (?)	
59–61	Clay, blue-gray, micaceous, contains some gravel.
61–68	Sand, white, medium-grained; contains a little gravel and a few clay particles.
68–74	Sand, gray-white, medium-grained.
74–75	Pea gravel.
75–92	Sandy clay, dark gray, micaceous, some brown clay particles.
92–108	Clay, dark blue-gray; some white gravel, carbonaceous matter and brown clay particles.
108–126	Clay, grades down into quartz pea gravel at base.
126–143	Sandy clay, gray, fine sand.
143–159	Blue clay.
159–171	Blue clay with streaks of fine sand.

171–212	Sand, gray, fine-grained.
212–213	Laminated sand and clay.
213–223	Sand, gray white, fine-grained.
223–271	Sand, fine- to medium-grained; with fossil fragments and clay particles.
271–290	Sandy clay, gray; with carbonaceous particles.
290–296	Peat bed.
296–307	Sand, gray, fine; with carbonaceous particles.
307–316	Sandy clay, gray; with carbonaceous particles.
316–344	Sand, gray, fine.

The base of the Pleistocene fill is placed between the red, yellow, and white subaerial beds and the blue to gray estuarine beds. The Pleistocene may, however, extend down to the base of the lowest gravel bed at 126 feet. It is also possible that the lowermost beds may be Miocene. In an earlier test well drilled at this same site a large number of water-worn *Rangias* were collected at 150 to 153 feet. These were submitted to Julia Gardner of the Section of Paleontology and Stratigraphy of the U. S. Geological Survey, who reported that the shells closely resembled *Rangias* from coquina of the Charlton formation (Pliocene) collected at the A. C. L. railroad bridge across St. Marys River, 3 miles southeast of Folkston, Charlton County, Georgia. She states (personal communication):

"Although the material, both from the Charlton formation and from the G. M. & O. water well, is poorly preserved, the *Rangias* in the two collections seem to be identical. The species in question is larger and heavier than *Rangia johnsoni* Dall but is more closely related to it and may be the Pliocene representative of that upper Miocene species."

Submerged valley of Mobile Bay.—The bottom of Mobile Bay is quite flat. Soundings by the U. S. Coast and Geodetic Survey show an average depth of 10 to 11 feet below mean low-tide level. Opposite the mouth of the bay and the western tip of Mobile Point is a submerged arcuate delta, which has a base about 10 miles wide and extends 4 miles out into the Gulf. Because the top of this submerged delta is also about 10 to 11 feet below mean low tide, it is evident that it is the delta of Mobile River and of the same age as the submerged valley of the river.

These features indicate that the most recent event in the Pleistocene history of coastal Alabama was the submergence up to present sea level of a river valley and delta formed during post-Wisconsin time. This recent rise in sea level was 10 to 11 feet.

Summary and Conclusions

After its deposition, the late Pliocene or early Pleistocene Citronelle formation was slightly entrenched by consequent streams and then was tilted to its present Gulfward surface slope of 5.5 feet per mile. The eroded surface of the Citronelle was then submerged by the sea to a level of 190 feet to 210 feet above present sea-level. The resulting embayed, compound shore line was eroded by the sea to form the Coharie terrace. Four other marine terraces, each outlining a compound shore line, occur at lower levels. The Sunderland terrace appears at 150 to 160 feet, the Wicomico terrace at 90 to 110 feet, the Penholoway terrace at 60 to 70 feet, and the Pamlico terrace at 20 to 30 feet. Several remnants of river terraces at the Coharie level occur in the upper Mobile River Valley. An extensive Mobile River terrace slopes down to the Penholoway level on the east side of the Mobile Valley. A third Mobile River terrace developed on the west side of the valley slopes down and grades into estuarine and marine terraces of Pamlico age. A submerged Mobile River floodplain and delta lies 10 to 11 feet below present sea-level and was formed during the post-Wisconsin.

The terraces in Alabama are named on the hypothesis that they correlate with Cooke's terraces in Florida and the Atlantic coast. No attempt is made to correlate them with Pleistocene stages.

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STRATIGRAPHY AND STRUCTURE OF THE OAK HILL SUCCESSION IN VERMONT

By Verne H. Booth

CONTENTS

	Page
Abstract	1131
Introduction	1132
Acknowledgments	1132
Previous work	1132
Physiography	1134
Stratigraphy	1135
Introduction	1135
General correlation	1136
Tibbit Hill schist	1138
Pinnacle formation	1141
White Brook dolomite	1145
West Sutton formation	1147
Gilman quartzite	1148
Dunham dolomite	1149
Parker slate	1151
Rugg Brook formation	1151
Skeels Corners slate	1153
Undifferentiated rocks of the eastern terrane	1153
Interpretation of the stratigraphic record	1155
Intrusives	1158
Metamorphism	1158
Structural features	1159
General setting	1159
Anticlinorium	1159
Homocline	1159
Thrust faults	1160
General statement	1160
Brigham Hill thrust	1160
Hinesburg thrust	1161
Arrowhead thrust	1162
Fairfield Pond thrust	1163
Klippen west of the Oak Hill escarpment	1164
Minor thrusts	1166
Mechanics of deformation	1166
References cited	1167

ILLUSTRATIONS

Figure	Page
1. Sketch map of northwestern Vermont and southern Quebec	1133
2. Regional geologic map of northwestern Vermont	1135

Plate	Facing page
1. Geologic map and structure sections of parts of the St. Albans, Enosburg Falls, Jay Peak, Milton, Mt. Mansfield and Burlington quadrangles, Vermont	1131
2. Structures	1144
3. Photomicrographs	1145

Abstract

North of the Winooski River, Vermont, the Oak Hill succession occupies the region between the eastern margin of the Champlain Lowland and the western foot of the Green Mountains. The lithologic units are traceable into southern Quebec, and some are traceable southward into west-central Vermont. Lower and middle Cambrian ages have been assigned to all units. The strata below the Gilman quartzite are tentatively correlated with the Mendon series in the Rutland-Brandon region. All units above the Gilman have stratigraphic equivalents in the adjoining Rosenberg succession.

The earliest recorded differential downwarping of the local geosynclinal area immediately preceded the beginning of deposition of the Pinnacle sediments. The distribution of the shaly, sandy, and conglomeratic facies of the Pinnacle and West Sutton formations indicate a western source for the sediments and also that the sea was encroaching westward. The Gilman quartzite marks the first occurrence of reworked sands; reworked sediments formed practically all of the clastics sediments thereafter.

A prominent anticlinorium occupies much of the southern part of the area. Northward it fades into a westward-dipping homocline. The western margin of the succession is formed in most places by a series of thrust faults, which may be continuous. The Oak Hill thrust, considered the western boundary of the succession in southern Quebec, has not been recognized in Vermont. The strata were strongly folded before thrusting; the thrusts are subsequent shear thrusts. Several klippen, related genetically to the thrusts, form prominent hills west of the western margin of the succession.

Introduction

This study, covers an area of about 300 square miles, forming an elongated triangular wedge whose apex is near the village of Williston in the Winooski River Valley about 7 miles east of Burlington Vermont, and whose base is that part of the international boundary which lies north of and between the villages of Franklin and Richford (Fig. 1). The western boundary is approximately coincident with the eastern edge of the Champlain Lowland. The eastern boundary is irregular and not everywhere clearly defined; it extends in a great crescentic curve from the vicinity of Williston to the international boundary at a point about 1½ miles northeast of Richford. The area includes a large part of the southern extension of a belt that Clark (1934, p. 9), in Quebec, termed the Oak Hill slice and described with other such thrust slices in southern Quebec. The structural relationships of these thrust slices are not so clear south of the international boundary; therefore, the rocks described are simply referred to as the Oak Hill succession and those beneath the Champlain Lowland to the west are referred to as the Rosenberg succession. Much of the Enosburg Falls quadrangle the eastern half of the Milton quadrangle, and smaller parts of the Jay Peak, St. Albans, Mt. Mansfield, and Burlington quadrangles are underlain by the Oak Hill succession.

During the summers of 1937, 1938, 1939, 1941, 1942, 1945, and 1946 about 38 weeks were spent in the field mapping the structure. To prove the areal continuity, the formations were walked out as far as possible. Extensive reconnaissance of outlying areas was undertaken. In much of the area, even in forested areas, outcrops are abundant, but the present topographic maps of these areas are too generalized in places for exact location. Structural complications, the lateral gradation of both primary and secondary facies, and the lack of continuous outcrop of certain key strata made location of formational contacts arbitrary in many areas. The structure and stratigraphy in several critical and highly significant areas were studied in detail, and the writer believes that he has obtained the facts necessary for the general structural picture. Dips and strikes were observed, but in this region dips and strikes may change as much as 90° in a few yards, or the dips may be very nearly vertical on both limbs of folds over large areas. In many places the dip and strike of the foliation is all that may be observed; these may or may not be the same as for the bedding.

Acknowledgments

Professor Marshall Kay of Columbia University has supervised the field work and directed the organization of this paper. His advice, interest, enthusiasm, and patience in the work have made this study possible. The writer is also indebted to Professor Walter H. Bucher of Columbia University and to Professor Marland Billings of Harvard University, each of whom spent several very instructive days with the writer in the field. The author expresses his appreciation to Dr. Wallace Cady of the United States Geological Survey and Professor Wilbur G. Valentine of Brooklyn College, both of whom spent considerable time in the field with the author and made many valuable suggestions. Doctor Cady and Professor T. H. Clark of McGill University read and criticized the manuscript. Professor Percival Raymond, Mr. Alan Shaw, both of Harvard University, and Mr. Solon Stone of Trinity College also made valuable suggestions.

Previous Work

In the Champlain Lowland just west of the area under consideration are the well-known Cambro-Ordovician successions, first studied in 1820 by Edwin James (1820, p. 250) and by now the subject of a voluminous literature. The two most recent discussions have been by Schuchert (1937) and Cady (1945). East of the area are the foothills and main ranges of the Green Mountains. Most writing on the geology of the Green Mountains is generalized and has been about that part south of the Winooski River. Almost nothing detailed has been written concerning the area between the Green Mountains and the Champlain Lowland in northern Vermont. Some general comments were made by Adams (1846), Hitchcock (1861), Keith (1923; 1932), and Longwell (1933). Jacobs (1938) published a section, based on reconnaissance studies, across the Milton and

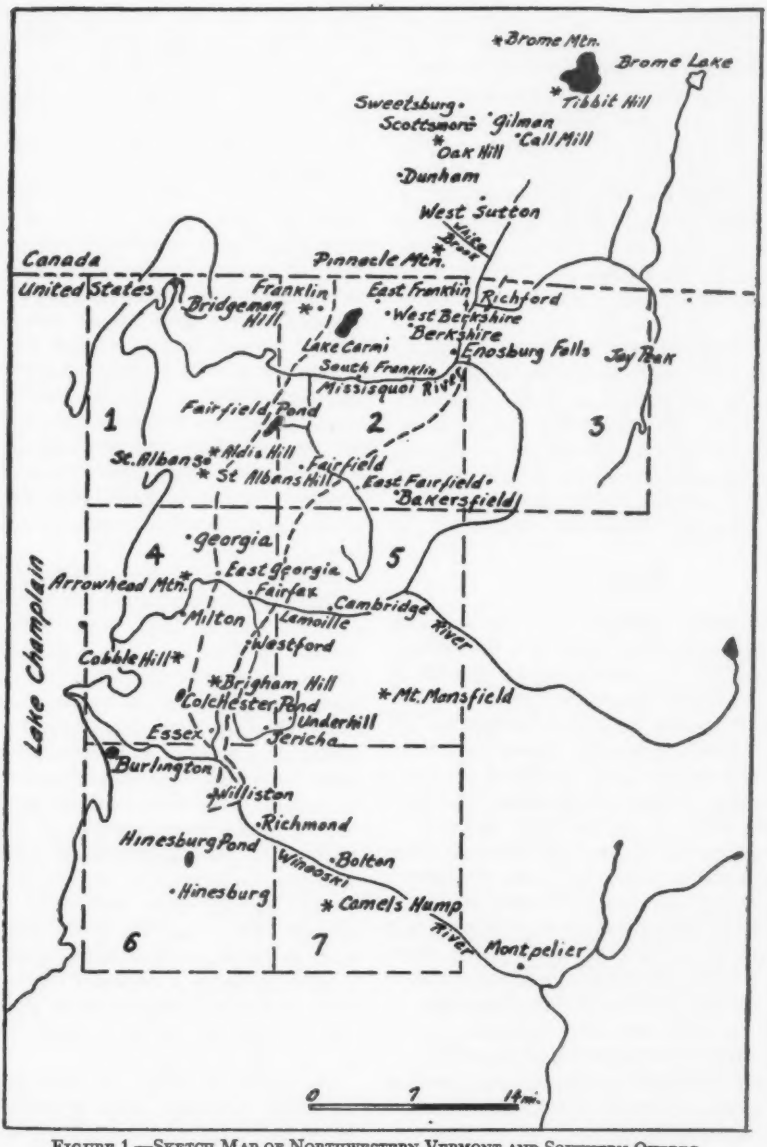

FIGURE 1.—SKETCH MAP OF NORTHWESTERN VERMONT AND SOUTHERN QUEBEC

– – – – Approximate boundary of area mapped in detail.

Quadrangles: 1. St. Albans; 2. Enosburg Falls; 3. Jay Peak; 4. Milton; 5. Mt. Mansfield; 6. Burlington; 7. Camels Hump.

Mt. Mansfield quadrangles. More detailed work has been done in the Province of Quebec, the type section of the Oak Hill succession, where T. H. Clark (1934; 1936) mapped the formations. Logan (1863) had mapped the formations much earlier, but his work had apparently

been forgotten (Clark, 1936, p. 136). J. B. Perry (1868, p. 344) had recognized what is now known as the Dunham dolomite in Franklin Township and traced it into southern Quebec nearly to the type locality as described by Clark (1936, p. 147). He correlated the dolomite with the "Red Sandrock Series" south of the Winooski River. Elkanah Billings (1872, p. 145) suggested that the marble beneath the slates at Swanton and St. Albans is the same as that which overlies the "quartz rock" (Cheshire or Gilman) of the Green Mountain Front.

The deplorable lack of interest in this sequence in Vermont, despite the enormous interest in the formations of the Champlain Lowland for over a hundred years, can be explained by the lack of fossils and any deposits of economic importance, the complexity of the structure, and the assumption that all formations below the Gilman were Precambrian.

Physiography

The region ranges between 300 and 1300 feet above sea level; much of it lies between 600 and 800 feet. This contrasts with 200 to 400 feet for the Champlain Lowland and 2000 to 4000 feet for the Green Mountains. The rise in elevation from the general level of the Lowland to that of the foothills of the Green Mountains is due to the increased resistance of the rocks of the Oak Hill succession to weathering and erosion. This increased resistance is due in part to the types of sediments deposited, in part to a gradual increase in metamorphism, and in part to the amount of vein quartz. The western margin is nearly everywhere a steep west-facing escarpment held up by the very resistant Gilman quartzite. In such regions as Arrowhead Mountain, St. Albans Hill, Aldis Hill, and Bridgeman Hill, the Gilman and higher formations outcrop west of the escarpment, henceforth referred to as the Oak Hill Escarpment.

The area is, in general, very hilly; many of the slopes are steep and rocky. In much of the area outcrops are very abundant; glacial activity has produced abundant roches moutonnees, but some critical bedrock areas are covered. Valleys are commonly developed along fault lines; this makes location of the extension of certain faults highly inferential.

Three major streams—the Winooski, the Lamoille, and the Missisquoi—cross the region; all rise east of the Green Mountains and flow generally westward to Lake Champlain. They are supposedly superimposed streams. Johnson's hypothesis (1931) of superposition from a Cretaceous coastal plain cover for the major streams flowing into the Atlantic south of the New England area creates the problem of a westward direction of flow for these streams. Geomorphologists have offered no solution. Most of the smaller streams follow the strike of the foliation of the rocks for all or a part of their courses. However, the resistance of some of the formations varies considerably along the strike; the streams cut across the strike for some distance and then resume courses roughly parallel to the strike of the foliation. The results are imperfectly developed trellis drainage patterns.

In general, the northern tributaries of the major streams are much shorter than the southern ones. None of the northern tributaries of the Missisquoi rise more than 4 miles from it, whereas its longest southern tributary, the Black River, rises within 3 miles of the Lamoille. Brown's Creek rises near Underhill Center, flows westward to Underhill where it is joined by a southward-flowing stream which rises 5 miles to the north, continues west-southwest to Jericho where it enters the valley of the Winooski, then shortly turns northward and flows into the Lamoille 11 miles to the north. Glacial damming might explain this unusual behavior. The explanation of the asymmetry of the drainage patterns is further complicated because the plunge of the major folds is southward instead of northward. Major cross folds are not located in the proper latitudes to serve as explanations.

The apparent subsequent drainage pattern is fairly well developed. Some of the northern tributaries of the Winooski and the southern tributaries of the Lamoille flow in continuations of the same valleys; the divides are imperceptible to the eye in a traverse along these valleys. The same is true for the northern tributaries of the Lamoille and the southern tributaries of the Missisquoi. Several southern tributaries of the

Winooski flow in continuations of the valleys of the various major tributaries of the White River, which, unlike the above three master streams, flows southeastward into the Connecticut River.

Stratigraphy

Introduction

The Oak Hill succession takes its name from Oak Hill (Fig. 1), about 8 miles north of Pinnacle Mountain in the Province of Quebec, where Gilman quartzite, Dunham dolomite, Oak Hill slate, Scottsmore quartzite and conglomerate, and Sweetsburg slate are exposed. The older formations are exposed a few miles to the south and southeast, on Pinnacle Mountain, in the headwaters of White Brook, and at West Sutton (Fig. 2). Clark (1934, p. 10) gave the stratigraphic succession in Quebec as follows:

	Thicknesses (Feet)
Lower Cambrian	
Vail slate	
Sweetsburg slate	
Scottsmore quartzite	300
Oak Hill slate	
Dunham dolomite	40–150
Gilman quartzite	2000
West Sutton slate	40–250
White Brook dolomite	20–75
Pinnacle graywacke	400
Call Mill slate	1–100
Precambrian:	
Tibbit Hill schist	Base not exposed.

Later (1936) he made the Vail slate a member of the Sweetsburg; he estimated the Gilman to be 3000+ feet thick; and he stated that the Tibbit Hill might very well be Cambrian.

The writer cannot identify all these formations in Vermont. The slate along the Oak Hill escarpment east of St. Albans Hill is probably the equivalent of the Sweetsburg slate, but the name Skeels Corners is used here since it can be definitely correlated with the Skeels Corners slate in the Rosenberg succession. Except for this area, the Skeels Corners is either absent from the Oak Hill succession or cannot be differentiated until more detailed work has been done on the slates in the Rosenberg succession. The Vail slate member is not recognized. Clark's Oak Hill slate is here termed the Parker slate, its correlative in the Rosenberg succession, to avoid multiplicity of names. Likewise, the term

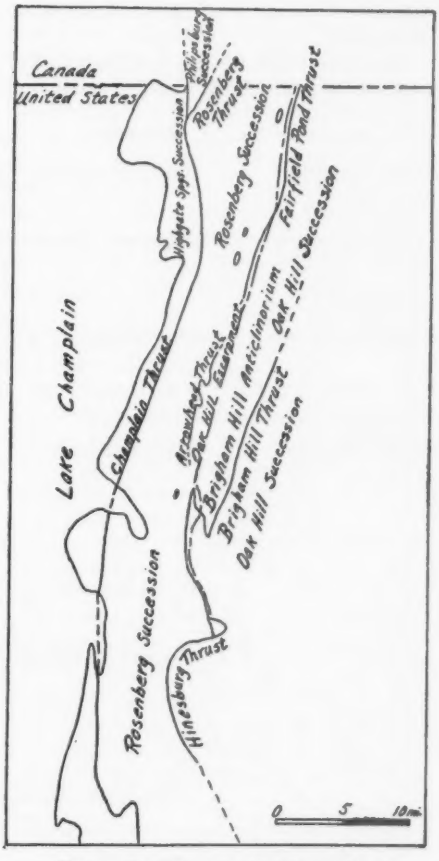

FIGURE 2.—REGIONAL GEOLOGIC MAP OF NORTHWESTERN VERMONT
Modified after Cady (1945, Fig. 5, p. 563)

Rugg Brook is used for the Scottsmore of Quebec, even though its differentiation from other formations in the Rosenberg succession is not always certain. The Call Mill slate has not been identified as a formation; if present, it is rarely over 1 or 2 feet thick. Clark (1936, p. 138) states that the Call Mill is seen neither at the international boundary nor to the north until about 2 miles south of West Sutton, probably because of nondeposition, for in several places

the Pinnacle graywacke lies directly upon the Tibbit Hill. If present in Vermont, it is included in the Pinnacle formation. The Pinnacle graywacke and the West Sutton slate are called formations because of the great variability of their rocks.

The stratigraphic succession of the Oak Hill sequence in northwestern Vermont is as follows:

Age	*Formation*	*Lithologic characteristics*	*Thickness (Feet)*
Middle Cambrian	Skeels Corners slate	Blackish gray slates with interbedded sandy layers and dolomitic lenses in places.	?
Lower Cambrian (?)	Rugg Brook formation	Buff- or brown-weathering sandy dolomite or dolomitic quartzite. The upper part contains limestone boulders in a sandy matrix.	0–40
Lower Cambrian	Parker slate	Blackish gray slates with interbedded sandy layers and dolomitic lenses in places. Sometimes limy, sometimes siliceous, sometimes pyritiferous.	25–100
Lower Cambrian	Dunham dolomite	Buff to chocolate/brown-weathering dolomite. Often sandy. Usually gray or grayish tan on fresh surface but may be pink, purple, tan, yellowish or white. Often contains siliceous or shaly interbeds.	40–300
	Gilman quartzite	White to gray-weathering quartzite, arkosic quartzite, argillaceous and slaty quartzite.	3000
	West Sutton formation	Typically a gray-black sandy slate or phyllite, but exceedingly variable. In places a graywacke, a quartzite, a pebble, or boulder conglomerate. Often hematiferous in its lower part.	200–400
	White Brook dolomite	Light to dark chocolate-brown weathering impure sandy dolomite. Color on fresh surface variable, light tan being most common. Often contains thin interbeds of slate, in places conglomeratic.	0–75
	Pinnacle formation	Lithologic characteristics variable. Typically a greenish-gray to light gray graywacke, often containing beds of sandy slates, black and green slates, quartzitic schists, and various types of conglomerates.	300–500
	Tibbit Hill schist	Typically a greenstone. Occasionally a light gray, blackish gray or greenish gray. Often amygdaloidal. In places large knots of epidote occur.	Base not exposed.

General Correlation

Clark (1934, p. 9; 1936, p. 136) designated the Oak Hill succession, with the possible exception of the Tibbit Hill schist, as Lower Cambrian. The Lower Cambrian age of the Gilman is established. The West Sutton, White Brook, and Pinnacle formations apparently

conformably underlie the Gilman and so may also be Lower Cambrian, although no fossils have been found. The metamorphism of these rocks is what would be expected in formations of their lithologic characteristics and structural setting conformably below the Gilman. Presumably the Tibbit Hill schist, a series of ancient lava flows, may also be placed in this generally conformable series.

These formations older than the Gilman can reasonably be correlated with the Mendon series, which underlies the Cheshire quartzite in west-central Vermont. Much of the Gilman is probably the equivalent of the brown (lower) part of the Cheshire south of Hinesburg (Cady, 1945, p. 526, 528). Strata correlated with the lowest part of the Gilman may be locally absent in west-central Vermont; in the stratigraphic column they may go at the unconformity between the Cheshire and the Mendon series described by Foye (1918, p. 85), and best seen above the Falls of Lana on Sucker Brook east of Lake Dunmore (Cady, 1945, Pl. 10). This unconformity cannot be considered a major systemic break since there is no evidence of such a break below the Gilman in the Oak Hill succession. Furthermore, a west-east traverse from the Champlain Lowland into the Green Mountains in the northern part of the Middlebury quadrangle reveals no evidence of a break between typical Cheshire quartzite and a series of slaty quartzitic rocks strongly suggestive of the rocks in the lower Gilman and those below it. The Moosalamoo phyllite (Keith, 1932, p. 395) is similar to the phyllitic facies of the West Sutton formation. The Forestdale marble (Keith, 1932, p. 394) is commonly purer, a little coarser-grained, and more uniform in color than the White Brook dolomite. It may represent a local variation in sedimentary environment. The Nickwacket formation (Keith, 1932, p. 394) corresponds fairly well to the graywacke facies of the Pinnacle. The chief difference appears to be one of metamorphic effects; the former contains biotite, the latter chlorite. Cady (personal communication, 1948) states that his 1948 work in the Green Mountains south of the Winooski River strongly supports, and may prove, these correlations.

Lower Cambrian fossils have been found in the Gilman and the Durham, chiefly outside the area of this report. Correlation of the Gilman with the Cheshire quartzite south of Hinesburg (11½ miles southeast of Burlington) along the Green Mountain Front in west-central and southwestern Vermont offers no difficulty because of the similarity in lithologic characteristics, stratigraphic position, physiographic expression, and fossils. Clark (1936, p. 146) states that about half a mile southwest of the cemetery in Scottsmore, Quebec, the Gilman contains a few poorly preserved fossils. The most important is the Lower Cambrian brachiopod, *Kutorgina*. Jacobs (1938, p. 20) states that a Lower Cambrian hyolithid was found on the western slope of Bald Hill in Westford Township, Milton quadrangle. Walcott (1888, p. 235) found *Olenellus* and *Hyolithes* in the Cheshire near Bennington, Vermont, and so established its Lower Cambrian age.[1]

The Dunham dolomite is well established as the correlative of at least the lower part of the Rutland dolomite west of the Green Mountain Front in west-central Vermont. South and east of St. Armand Centre in Quebec Clark (1936, p. 147) has found fragmentary trilobites belonging to the Mesonacidae. This family is definitely Lower Cambrian. Keith (1923, p. 28) and Jacobs (1938, p. 21) refer to the Dunham as the Rutland. Cady (1945, p. 529) correlates the Dunhan of west-central Vermont with the lower Rutland. He also (p. 526) recognizes the Dunham (Winooski, in part, of Keith) in the Rosenberg succession in the Champlain Lowland in northwestern Vermont; the correlation is made because the Dunham of the Oak Hill

[1] If one adopts the base of the biozone of *Olenellus* as the base of the Lower Cambrian, as suggested by Wheeler (1947, p. 157), then much of the Gilman, and probably much of the Cheshire, and all of the formations below it, would have to be called pre-cambrian. All the fossils found appear to be from the upper parts of the Gilman and the Cheshire. In formations where fossils are found as sparingly as in these, one may infer that conditions for preservation were not at all favorable, that the very few preserved are "accidents." *Olenellus* may have lived long before one was preserved. It may be desirable to always use the lowermost known occurrence of *Olenellus* as the base of the Cambrian in order to get away from diastrophism as a time index and to fix a boundary line between the Cambrian and the Pre-cambrian on the same basis as other period boundary lines, but this does not necessarily make for more precise correlation as to time of sedimentation.

succession directly to the east is strikingly similar lithologically, and the rocks in each area occur in similar positions in comparable stratigraphic successions. The Mallett member (Cady, 1945, p. 528) is either absent or not sufficiently important or distinctive to be mapped as a unit in the Oak Hill succession.

Fossils have not been reported from the formations above the Dunham (except in the Skeels Corners slate) in the Oak Hill succession. Stratigraphically and lithologically the Oak Hill slate of Clark appears to be the equivalent of the Parker slate in the Rosenberg succession. Lower Cambrian fossils, chiefly trilobites, have been found in the Parker of the Rosenberg succession in areas where there is little cleavage (Schuchert, 1937, p. 1026); none has been found in the well-cleaved Parker of the Oak Hill succession. There are bioherms in the Parker of the Rosenberg succession (Schuchert, 1937, p. 1025); none has been found in the Parker of the Oak Hill succession, but limestone breccias in the succeeding formation, the Rugg Brook, suggest their presence in the original area of deposition of the Parker. In both successions the Parker has disconformities both above and below it (Schuchert, 1937, p. 1025, 1026, 1034; Clark, 1936, p. 148). Cady (1945, p. 526, 531) places the Parker of the Rosenberg succession above the Dunham; this is the position of the Oak Hill slate in the Oak Hill succession in Quebec. Shaw (1948, personal communication) believes these slates in the two successions are the same.

The Rugg Brook formation resembles, in one or another of its facies, several of the Cambrian sandy dolomites and breccias in the Rosenberg succession. The stratigraphic position of this formation in the Oak Hill succession, above slates that lie on the Dunham and below the Skeels Corners slate, strongly suggests that it is the same as the Rugg Brook in the Rosenberg succession. The Rugg Brook in Vermont is correlated with the Scottsmore (Clark, 1934) formation of Quebec on the basis of stratigraphic position and lithologic characteristics. The Rugg Brook in the Rosenberg succession is late early Cambrian or early middle Cambrian (Schuchert, 1933, p. 366; 1937, p. 1039; Howell, 1934, p. 334; 1938, p. 97; Cady, 1945, Fig. 3, p. 522). Determination of the exact age of the Rugg Brook must await a comprehensive study of the Rosenberg succession in northern Vermont. Without additional data, Clark's Lower Cambrian correlation (1934, p. 6) of the Scottsmore (= Rugg Brook) stands.

The Skeels Corners slate of the Oak Hill succession is correlated with the Skeels Corners slate of the Rosenberg succession on the basis of two specimens of *Bovicornellum vermontense* Howell. One was found by Alan Shaw (1948, personal communication) along the base of Oak Hill escarpment east of St. Albans Hill, the other by the writer in the same area. *Bovicornellum vermontense* is a peculiar hornlike guide fossil for the Upper Cambrian according to Schuchert (1937, p. 1040), but Howell (1939, p. 1964) states that it is either late Middle Cambrian or very early Upper Cambrian; Percy Raymond (1948, personal communication) believes it to be definitely Middle Cambrian on the basis of associated trilobites. It is found in the typical Skeels Corners (Howell, 1939, p. 1964). The correlation of the Skeels Corners is complicated by the presence of several similar slates in the Rosenberg succession and by several disconformities above the Dunham, already referred to in the discussion of the Parker. Howell (1939, p. 1964, 1965) lists five disconformities above the middle Cambrian St. Albans shale. The correlation of the Skeels Corners with the Sweetsburg slate of Quebec rests solely on the fact that both lie above the Scottsmore formation.

Tibbit Hill Schist

The type locality of the Tibbit Hill schist is on Tibbit Hill in Quebec (Clark, 1934), west of the southern end of Brome Lake and 15 miles northeast of Pinnacle Mountain (Fig. 1). The schist is typically a greenstone.

MEGASCOPIC FEATURES: Most commonly it is dark green; variations are yellowish green, bright green, light gray, medium gray and blackish gray. It weathers dark greenish or grayish black. It is most commonly fine-textured; some facies are medium coarse-textured, with the larger grains of magnetite, epidote, and actinolite barely visible to the naked eye. Parts of the schist are vesicular and amygdaloidal; one medium- to light-gray variety is amygdaloidal. The amygdules range from those barely visible to some over an inch long. In

places, particularly in the eastern part of the main belt of outcrop, the amygdules are drawn out parallel to the foliation. They consist largely of quartz, albite, epidote, and calcite. Chlorite and perthite are occasionally seen. Many of them are composed of one mineral; some are mixtures in various combinations. Much of the albite is obscured by aggregates of kaolin and sericite. Occasionally an amygdule occurs with an epidote core surrounded successively by albite and chlorite. Knots of epidote 1 to 6 inches across are striking in many outcrops. Knots of epidote 1–2 feet across are associated with considerable quartz, and sometimes chlorite.

West of Ayers Hill, south of the international boundary and on the extreme western edge of the Jay Peak quadrangle, one small area contains circular masses of epidote and quartz in the form of "pots" or "bombs" in the Tibbit Hill and ranging from 5 or 6 inches to 15 inches in diameter. Only a very few show any part of the vertical dimension. One of these measured 15 inches across at the top, tapered downward, and then expanded again. The exposed part was 26 inches long. All are upright, their longest dimensions essentially parallel to the foliation of the enclosing rock. None shows any foliation. They consist chiefly of an intimate mixture of acidic feldspar and epidote with occasional grains of chlorite; the enclosing rock consists chiefly of acidic feldspar and chlorite with occasional grains of epidote. The physical contrast with the enclosing rock is more striking than the difference in composition would suggest. Similar masses of epidote and feldspar, except that they are tabular and parallel to the foliation, occur in the same area. None are over a few yards long, and they may be up to 1–2 feet wide. No adequate explanation of the origin of these forms can be given.

Microscopic Features: Essential minerals of the schist are epidote, albite, and chlorite. Almost all the albite grains are small; many are somewhat irregular and more or less equidimensional. Twinning is not pronounced. Epidote is sparingly present in some parts of the schist, but in others it forms the great bulk of the rock. Generally it is anhedral to subhedral but occasionally euhedral. The chlorite is also unevenly distributed throughout the formation. Normal and abnormal interference colors may appear in the same section. An increase in albite generally means a greater decrease in epidote than in chlorite. In a few specimens from the eastern part of its main belt of outcrop where metamorphic effects are greater, actinolite is an essential mineral.

The chief of many minor constituents are magnetite, magnetite-ilmenite, carbonate, sphene, and leucoxene. Others are sericite, quartz, apatite, tourmaline, hematite, biotite, bleached biotite, rutile, and zeolites. In some specimens magnetite is so abundant that it might well be classed as an essential mineral. Usually it occurs in small irregular grains throughout the rock; many of the grains are elongated in the direction of the foliation. The partial or complete alteration of metallic grains to leucoxene, common in some specimens but absent in others, suggests a titaniferous magnetite along with the ordinary magnetite. Carbonate occurs in most specimens in irregular patches; in some it is abundant. Traces of malachite and chalcopyrite occur in one area, about 2½ miles north of the village of Berkshire. An attempt was once made to mine them.

Metamorphic Effects: The Tibbit Hill is essentially a schistose rock, but the variability of this structure is pronounced. The medium- to light-gray, always amygdaloidal facies usually lacks schistosity except in the eastern part of the area of outcrop; in some phases the schistosity is obscure, and in others it is extremely well developed. The development of the foliation tends to increase eastward, as do other metamorphic features. The highly schistose facies in association with the more massive and amygdaloidal facies can be seen excellently in the valley of the unnamed brook that is crossed by the road from Ayers Hill (extreme northwestern corner of the Jay Peak quadrangle) to Richford. Here parts of the outcrop are more properly slates and phyllites rather than schists, very different megascopically from the typical Tibbit Hill. These schistose beds may have originally been tuffs. In some areas gray lenslike stringers and small rounded schistose masses form bands which appear to be interbedded with the typical greenish facies. Thin sections show them to be essentially the same as the typical rock. These may be observed along the road from Enosburg Falls to West Enos-

burg about half a mile from the bridge across the Missisquoi.

INFERRED PRIMARY CHARACTERS: The vesicular and amygdaloidal character of parts of the schist attests that it has been derived from a series of lava flows. Some specimens show suggestions of original flow structure. The schistosity almost always masks the flow structure; rarely, a structure interpreted as due to flow was at a small angle to the schistosity. The high percentage of chlorite, epidote, and magnetite in the schist suggests a basic flow. The presence of considerable albite and epidote and the absence of any other feldspar indicate a nearly complete decalcification of the original feldspar (presumably labradorite or andesine). No trace of a pyroxene or of an amphibole, other than actinolite in a few specimens, has been seen. The only other ferromagnesian mineral observed in the 30 thin sections studied was biotite in two or three of them. The primary ferromagnesian mineral, probably a pyroxene, is apparently altered to chlorite, epidote, carbonate, and magnetite. The actinolite in the coarser-grained facies is probably not original; it may have been formed by the dynamothermal metamorphism of the primary ferromagnesian mineral or of its alteration products. The same metamorphism produced the coarser grain and the schistosity of the rocks. These actinolite-bearing rocks are found in the eastern portion of the area, nearer the axis of the Green Mountains, and are associated with rocks that represent a higher grade of metamorphism than the rocks of the western portion of the Oak Hill terrane.

The rock is at present essentially an albite-epidote-chlorite schist. According to Harker (1939, p. 280) this is the type of rock expectable when basic igneous rocks are regionally (dynamothermally) metamorphosed. However, there is some evidence of other factors in its formation. The amygdules, the conspicuous knots of epidote, the "pots" and "bombs" in the Ayers Hill area, the tabular masses of acidic feldspar and epidote, and the tiny veins containing chlorite, epidote, carbonate, and quartz cannot be explained by dynamothermal metamorphism. These may have been produced by the residual liquors of the magma working their way through the vesicles and fissures of the lava. The same liquors are capable of altering the ferromagnesian minerals and of decalcifying the feldspars. Such changes in an igneous rock, termed autometamorphism (Tyrrell, 1929, p. 329), are on the borderline between truly igneous and truly metamorphic effects. The albitization of basic igneous rocks is a widespread phenomenon of autometamorphism (Tyrrell, 1929, p. 329). The writer cannot separate entirely the effects of autometamorphism from the effects of dynamothermal metamorphism.

It is inferred that the original rock was an olivine-free basalt or a less basic rock. Harker (1939, p. 287) states that igneous rocks of intermediate acidity, subjected to dynamothermal metamorphism, produce an albite-epidote-sericite-chlorite schist rather than an albite-epidote-chlorite schist (as is produced by such metamorphism of olivine-free basalts). The degree of metamorphism in both cases is moderately low grade. The former has more sericite and albite and less chlorite and epidote than the latter. In the thin sections examined, in some specimens the albite is by far the most abundant mineral, some sericite is present, and epidote is almost lacking. In others epidote is by far the most abundant mineral, some albite is always present, and sericite is lacking. In still others albite and epidote are about equal. Chlorite appears in all sections but is most abundant in those in which albite is the chief mineral. These variations are explained by assuming a series of lava flows which varied in basicity. The thickness of the schist appears too great for a single flow. Also, interbedded with the lavas are some thin beds of detrital origin that contain the same minerals as do the lavas but carry more sericite and less albite. A small amount of quartz is also present. These detrital rocks are extremely fine-grained and very schistose. Thus there may have originally been two or more lava flows, separated in time long enough for a regolith to develop. Part of the regolith was eroded before the next flow. Later the higher formations of the Oak Hill succession were deposited, and then the whole was dynamothermally metamorphosed.

DISTRIBUTION: The width of outcrop of the Tibbit Hill schist along the international boundary reaches a maximum of 4½ miles. It extends northward into Canada for more than 20 miles and maintains most of this width for much of that distance. Southward toward the Missisquoi

River, it disappears beneath the younger Pinnacle formation except in the eastern part of its belt of outcrop. Here it continues for a few miles farther south before disappearing beneath the Pinnacle. Farther south it reappears in numerous anticlinal inliers as far as the Lamoille River, and possibly beyond. Four to six miles southeast of St. Albans an isolated belt of a similar rock has been mapped as Tibbit Hill on the basis of its megascopic appearance, its obvious igneous character (as determined from study of thin sections), and its stratigraphic position. It lacks the epidote so characteristic of the typical Tibbit Hill, but this may be because metamorphic effects are somewhat less to the west.

The thickness of the Tibbit Hill is unknown, since its base is nowhere exposed. It must be several hundred feet thick in order to be exposed over such a large area in a region of nearly 1000 feet of relief. It is either very early Cambrian or late Precambrian.

Pinnacle Formation

Clark (1934) named this formation the Pinnacle graywacke from its occurrence on the top of Pinnacle Mountain in southern Quebec (Fig. 1). The extreme variability of texture and composition, especially south of the Missisquoi River, makes designation by a single rock type misleading; some parts are more arkosic than graywackish. The usage of the term graywacke needs reviewing, for Krynine (1941) states, "the common mistake of confusing arkoses (rich in feldspar) with graywackes (rich in non-feldspathic rock fragments) is sure to lead to petrologic and interpretive disaster."

Graywacke has been defined in various ways. Rosenbusch (1923) states that graywackes are sandy Paleozoic sediments, generally gray, greenish gray, or brownish gray, less commonly brown or blackish; composed of very variable quantities of quartz, orthoclase, and plagioclase grains along with rounded or angular fragments of schist (= slate); quartzite, less commonly diabase, granite, and other rocks. The cement is scanty and is composed of clay-slate material (when unaltered); but carbonate and quartz cements are also known. In other cases augite, hornblende, garnet, tourmaline, apatite, and fragments of limestone are abundant too. The range of chemical composition is very wide. Krynine (1943) defines a graywacke as "a variety of sandstone made up of quartz, clay and rock fragments of phyllites, slates, schists, quartzite, chert, mica, etc." He states that popularly it may be defined as a "dirty sand full of clay and mica" for the normal graywacke, or as a "pepper and salt sand" for the strongly cherty varieties. Pettijohn (1943, p. 941–943) lists a number of other definitions of graywacke, with most of which he is in complete disagreement. He (1943, p. 941–957, 963–968; 1949, p. 243–257) gives a comprehensive discussion of graywacke, their formation, and the distinction between graywackes and arkoses, and states (1949, p. 252) that the origin of graywackes is only partially understood. He defines (1943, p. 943; 1949, p. 251) graywacke as a type of sandstone both uniform and distinctive in appearance, dark, generally tough and indurated, with large grains of extremely angular quartz (often sliverlike) and feldspar set in a prominent to dominant "clay" matrix which may on low-grade metamorphism be converted to chlorite and sericite and partially replaced by carbonate, usually ankerite. The paste of many graywackes equals or exceeds the volume of large detrital fragments. Present also are quantities of rock fragments, mainly chert, quartzite, slate, or phyllite. Graded bedding, intraformational conglomerates of shale or slate chips, slip-bedding, and an association with tuffs and spilitic greenstones are also characteristic. He also reports general absence of mineral cement; cross-lamination is rare, and ripple marks almost never are found.

He (1949, p. 258) describes a typical arkose as consisting primarily of quartz and feldspar (80–98 per cent). Potash- and soda-rich feldspars dominate. The cement is usually calcite or iron oxide. Silica is rare, and secondary outgrowths are very rare. Arkoses are coarse, are but moderately well sorted, and the grains are angular to subrounded. Of especial significance (p. 260–261) is the mingling of clear, wholly fresh feldspar with clouded, partially weathered feldspar. Arkoses may be strongly cross-laminated. Krynine (1940, p. 50) defines arkose as a highly feldspathic (30 per cent of feldspar or more) sediment derived from a granite and having the appearance of a granite.

The writer is inclined to accept Pettijohn's

concepts of graywackes and arkoses. The coarse-grained facies of the Pinnacle formation is thus more nearly a graywacke than an arkose. Typical graywacke features include angularity of the larger fragments, character of the matrix, proportion of matrix to larger fragments, common occurrence of carbonate, variability in the bedding, graded bedding, poor sorting, dark color, upward transition into rocks that are best described as subgraywackes and into slates and phyllites. Atypical features include the infrequent occurrence of typical arkose, feldspar in abnormal amounts in some areas, sericite in far greater abundance than chlorite in the matrix, strong cross-lamination (Pl. 2, fig. 1) in several outcrops, abundance of conglomeratic interbeds, scarcity of lithic fragments and chert, and lack of association with submarine pillow lavas of the spilitic suite. However, the occurrence of the atypical features is most common in the western outcrops, nearest the ancient shore line where shallow-water characters would be expectable.

MEGASCOPIC FEATURES: A coarse greenish-gray facies is predominant in most areas in the northern and western parts of the outcrop belt of the Pinnacle formation. At the southern base of the prominent hill about a mile north of the village of West Berkshire it is a light-gray arkosic rock; in a few areas it is a nearly pure quartzite. In many areas it is a pebble conglomerate, and in one area a boulder conglomerate. In this respect it differs from the Pinnacle of Quebec, which is never conglomeratic (Clark, 1936, p. 142). Locally the top part is a fine-grained black to gray slate or a coarse-grained sandy slate. Slates are commonly interbedded with siltstones, graywackes, phyllites, and, occasionally, with thin dolomitic lenses; in places these interbedded types make up the bulk of the formation. In a few small areas interbeds are composed chiefly of magnetite-ilmenite. Clark (1936, p. 142) states,

> "... In composition it varies from an apparently pure quartz-feldspar (chiefly orthoclase) combination through all gradations to a rock in which magnetite-ilmenite grains predominate to the virtual exclusion of all else."

Schistosity is poorly developed in the coarsest-grained facies but is always present in the medium- to fine-grained facies. The coarser-grained facies are commonly well bedded; the beds range from a fraction of an inch to 10 or more feet in thickness. In places it is strongly cross-bedded (Pl. 2, fig. 1).

MICROSCOPIC FEATURES: Quartz, feldspars, magnetite or magnetite-ilmenite, chlorite, and sericite are common in the coarse-grained facies. The feldspar is chiefly orthoclase; minor amounts of microcline, sodic plagioclase, and perthite occur. Estimates of feldspar from 5 to 40 per cent (rare) were made from thin-section studies. The proportion of quartz to feldspar also varies greatly, but quartz is almost always considerably more abundant. Most quartz and feldspar grains show strain. Some quartz grains are sliverlike. The large angular grains of quartz and feldspar are set in a matrix of these same minerals, plus sericite and chlorite. Sericite is far more abundant than chlorite. The proportion of matrix to large fragments varies within wide limits, but in this facies the former is less than the latter. Nearly always there is a liberal sprinkling of magnetite or titaniferous magnetite (Pl. 3, figs. 5, 6) apparently detrital. In a very few small areas, some beds contain so much of these metallics that specimens are an iridescent black and noticeably heavy. There are, however, no such concentrations of it as are found in some areas in Quebec (Clark, 1936, p. 142). Despite the abundance of these metallics there is little evidence of limonitic staining in the coarse-grained facies.

The finer-grained facies has a similar mineralogy except for the proportions; feldspar grains are relatively rare, and sericite is much more abundant. There are all gradations between the coarse-grained facies and the finer slaty and phyllitic facies which consist almost wholly of sericite, quartz, chlorite, and magnetite.

Tourmaline, zircon, sphene, pyrite, apatite, hornblende, rutile, leucoxene, kaolinite, epidote, biotite, and muscovite are present in small amounts. The pyrite, locally abundant, is an introduced mineral. Lithic fragments are never abundant in any specimen studied; many contain none. They are granite, aplite, slate, quartzite, and basic volcanics.

THE CONGLOMERATES: The conglomerates of the Pinnacle formation occur south of the Missisquoi River. All but one—a boulder conglomerate—are pebble conglomerates. In some, the

pebbles make up the bulk of the rock; in most they occur sparingly and unevenly scattered throughout certain beds. Most of these conglomeratic beds are not more than 1–2 feet thick; rarely are they over 10 feet. The stratigraphic position of the conglomerates within the Pinnacle was not determined because of much repetition by folding.

The boulder conglomerate is found in the northeastern corner of the Milton quadrangle along the Fairfax-St. Albans road slightly over a mile north of the former site of School No. 3. Here in shallow road cuts is exposed a pebble and cobble conglomerate, and in the pasture to the east and northeast is a succession of fine exposures of the boulder conglomerate. These extend half way down the eastern slope of the northern hill. Other outcrops occur not far west of the road. The boulders are all well rounded and range from a fraction of an inch to about 15 inches across. Most are of igneous origin, some are of slate, and a few are of arkosic and shaly quartzite or siltstone. Quartz pebbles are very rare; those present are small.

Microscopic examination of about 40 igneous boulders reveal that they consist largely of fine-grained to moderately coarse-grained syenites, quartz syenites, sodic granites, and microcline granites. While there is some variation among each of the above types, the essential composition of each type is as follows:

Microcline granites—microcline, albite-oligoclase, perthite, microcline-perthite, myrmekite, quartz.
Sodic granites —Albite-oligoclase, perthite, quartz.
Syenites and Quartz syenites —Orthoclase, albite-oligoclase, perthite, a little quartz.

Occasional pebbles or small boulders of finer-grained igneous rocks consist of trachytic porphyries, aplites, and a rock resembling a bostonite. Presumably these were from small intrusives, probably dikes.

Accessory minerals include zircon, apatite, sphene, tourmaline, and possibly magnetite. They are common to all the above types. Secondary minerals are chlorite, quartz, sericite, bleached biotite, carbonates, epidote, sagenite, kaolinite, hematite, pyrite, limonite, magnetite, titaniferous magnetite, and leucoxene. Ferromagnesian minerals, originally sparingly present, are completely altered to aggregates of chlorite, sericite, bleached biotite, calcite, hematite, magnetite, and limonite; not all are present in every aggregate. The primary minerals were probably hornblende and biotite. The alteration was hydrothermal and appears to have been highly selective, for the feldspars are almost unaltered. The rare grains of pyrite are always altered to limonite around the edges, and, in some cases, later metamorphism appears to have altered the limonite to hematite. Veinlets of quartz are common in the boulders; occasional veinlets of chlorite, epidote, carbonate, and pyrite are present. Some of the sections show carbonate, not in veinlets, and much too abundant to be formed entirely from the alteration of ferromagnesian minerals containing calcium. It is probably replacement.

Except for the microcline granites and the trachytic porphyries all these rocks are considerably darker than is expected of such acidic rocks; most are the color of basalts or gabbros, because of the abundance of magnetite and titaniferous magnetite that has been introduced into them (Pl. 3, figs. 1, 2, 3). These metallics are commonly found along grain boundaries, along the cleavage planes of the feldspars, and in fractures. They are so abundant in some specimens that they must have in part replaced other minerals. In some sections they are euhedral to subhedral, but most of them are anhedral. Skeletal crystals of titaniferous magnetite are common in some sections; the empty spaces are commonly bordered by leucoxene, and tiny streaks of leucoxene cross them. Many of the skeletal crystals are elongate and pointed at both ends (Pl. 3, fig. 3). Some short fine needles of leucoxene have migrated out into the adjacent minerals. Many smaller metallic grains in the same sections—presumably magnetite grains—are not altered. In rare cases the magnetite is associated with a little hematite. The magnetite in the boulders suggests the source of the magnetite grains that are so abundant throughout the Pinnacle and West Sutton formations.

All the igneous boulders show some dynamometamorphic effects. Quartz and feldspar grains show strain shadows; bent twinning in the feldspars is common; mortar structure sometimes appears about the quartz grains;

and in some specimens the margins of the feldspar grains have been crushed and the crushed material altered largely to sericite. The metallic grains show no effects of crushing; the metallics apparently were introduced after the release of the stresses which caused the dynamo-metamorphic effects.

Boulders of sedimentary origin are also present in the conglomerate. Slate is the most common, but they are extremely local in their occurrence. They are usually elliptical, rarely more than an inch thick, and up to a foot long. They are typical blackish-gray, fine-grained, well-cleaved slates. Only where they are in close contact with igneous or other resistant boulders do they show effects of crushing. Those in contact with resistant boulders are bent around them as if the slate were plastic. These slate boulders tend to be roughly aligned in the direction of the cleavage of the enclosing matrix. The igneous boulders have probably much interfered with their alignment. Other boulders are of siltstone, arkose, and impure quartzite. The mineralogy is about the same as for the granites except that there is less feldspar and more sericite. Some boulders are gneissoid due to the arrangement of the introduced metallics in streaks; this is more apparent in the hand specimen than in thin section.

The conglomeratic lenses are not confined to a single zone; they alternate chiefly with beds of coarse graywacke. The observable thicknesses of the boulder lenses vary from 2 to about 10 feet. None of the lenses have any great continuity along the strike; the maximum is a few hundred yards. Only one other area of such boulder conglomerates has been located; these occur in the West Sutton formation. Pebbles of similar igneous rocks occur in numerous regions in the Pinnacle formation; the pebbles are never over an inch in diameter, and they are sparingly distributed throughout the matrix. One such area is the inlier of Pinnacle about a mile southeast of the village of Sheldon (southwestern part of the Enosburg Falls quadrangle).

Some of the other pebble conglomerates are composed chiefly of quartz pebbles; in others the pebbles are claystone, slate, or siltstone. In other areas the pebbles consist chiefly of ironstone (limonitic claystone) pebbles associated with a few quartz and siltstone pebbles. These are usually embedded in a matrix of coarse angular grains of quartz. One such area is the south end of the 1000-foot hill about a mile east of School No. 9 in the southeastern part of the St. Albans quadrangle. In some areas the pebbles make up the great bulk of the rock; in others they are sparingly dispersed throughout the matrix.

The typical graywacke facies of the Pinnacle formation is best developed in the area north of the Missisquoi River and for a short distance southward. The relatively narrow strip of it at the international boundary expands greatly southward; to the north, erosion has cut relatively deeper into the formation because of the general southward plunge of the major structure. In this area the coarse-grained facies forms more than 90 per cent of the formation, though it never becomes truly conglomeratic. In a few places, slaty beds 1 or 2 feet thick separate the Tibbit Hill and the Pinnacle. These beds may be the equivalent of Clark's (1936, p. 138) Call Mill slate; they are here included in the Pinnacle because they do not form a mappable unit. In the north, the transition from coarse-grained graywacke to the White Brook dolomite is, in general, abrupt; intervening beds of finer-grained clastics increase toward the Missisquoi River. Outliers of Pinnacle with a synclinal structure occur in the Tibbit Hill schist; conversely, inliers of Tibbit Hill with an anticlinal structure occur in the Pinnacle. Only the larger inliers and outliers are mapped (Pl. 1). West of the main belt of outcrop of the White Brook dolomite south of the Missisquoi, the

PLATE 2.—STRUCTURES

FIGURE 1.—CROSS LAMINATION IN PINNACLE FORMATION[1]
1½ miles northeast of East Georgia.
FIGURE 2.—FOLD OVERTURNED TO WEST IN SKEELS CORNERS SLATE
¾ mile east of St. Albans Hill.
FIGURE 3.—LIMESTONE BRECCIA IN RUGG BROOK FORMATION
On Knoll on northeast side of Bridgeman Hill.

FIGURE 1

FIGURE 2

FIGURE 3

STRUCTURES

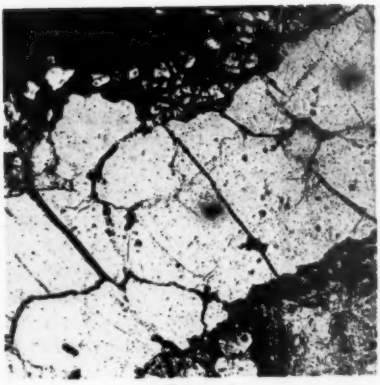

FIGURE 1

FIGURE 2

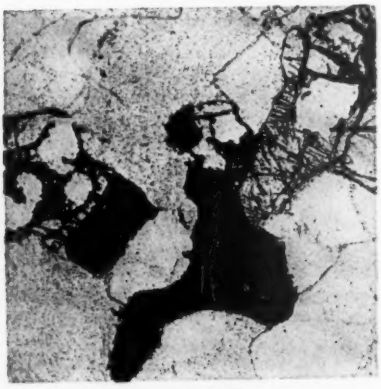

FIGURE 3

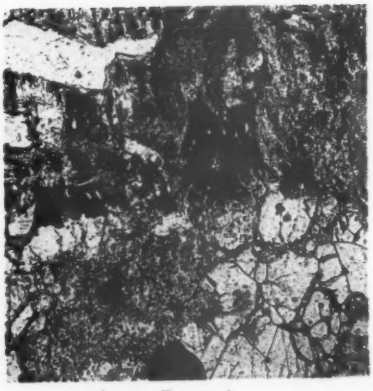

FIGURE 4

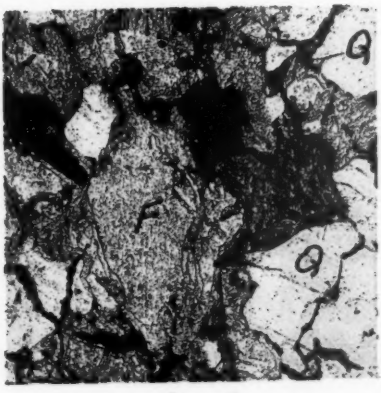

FIGURE 5

FIGURE 6

PHOTOMICROGRAPHS

Pinnacle appears as inliers, whose structure is necessarily anticlinal, in the West Sutton formation. The westernmost anticlinal inliers are of the typical graywacke facies which in places are sparingly conglomeratic, but to the east there is a gradual change to the finer-grained facies, especially in the top part of the formation. This change is best observed in the amphitheater developed by erosion on the north side of the prominent but small hill about three-fourths of a mile southwest of Pumpkin Village School and 3 miles northeast of the village of Fairfield. Here the Pinnacle consists of alternating beds of graywacke and slate; the beds range from half an inch to several inches thick. Still farther south near Fairfield village are small anticlines cored by Pinnacle slates which are distinguishable from the younger West Sutton slates surrounding the anticlines only because the White Brook dolomite intervenes. South of the Missisquoi River and east of the main belt of outcrop of the White Brook dolomite, the coarse-grained facies of the Pinnacle gives way somewhat abruptly along a highly irregular line to finer-grained slates, phyllites, schists, and graywackes.

Southwest of the village of Fairfield where the main belt of the White Brook takes a wide swing to the west, the coarse-grained facies—which here has numerous thin beds of pebble conglomerate—comes in again and continues southward through Fairfax, Westford, and beyond. West of the Brigham Hill fault and just north of the LaMoille River a large anticlinal inlier has been so deeply eroded that the Tibbit Hill schist is exposed in the core. Coarse-grained graywacke with some conglomeratic interbeds make up the largest part of this inlier. A short distance north of the LaMoille River, the Pinnacle and other formations forming this inlier plunge beneath the Gilman and are not known to the south. The top part of the Pinnacle is commonly a black to gray or brownish slate which in some outcrops is sandy. The transition from coarse Pinnacle to White Brook dolomite is abrupt in many areas.

Clark (1936, p. 141) states that the Pinnacle in its type locality is about 400 feet thick. The writer believes this is a good estimate of the average thickness, particularly in northern Vermont, although he could find no place to measure it accurately. About half a mile south of the international boundary and directly north of the village of West Berkshire, an estimate of 350 feet was made. Farther south and easterly, it may be thicker, but repetition by folding makes an estimate unreliable.

White Brook Dolomite

The type section of the White Brook dolomite (Clark, 1934) is in the region of the headwaters of White Brook about 6 miles northeast of Pinnacle Mountain in Quebec (Fig. 1). Typically an impure, variegated, finely crystalline dolomite, on fresh surfaces it is most often tan to whitish, but shades of gray are common, and in many places it is mottled with pink, violet, purple, or white streaks and patches. All colors may occur in the same outcrop. The streaks and patches are chiefly calcite; weathering does not change their colors. The great bulk of the dolomite weathers to some shade of brown; often it is a rich light chocolate brown easily visible from considerable distances. In one area, there are many small patches or nodules of

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PLATE 3.—PHOTOMICROGRAPHS

FIGURE 1.—PORTION OF GRANITE BOULDER IN PINNACLE FORMATION
Showing magnetite fractures in feldspar and along grain boundaries. × 37

FIGURE 2.—MAGNETITE GRAINS IN GRANITE BOULDER in Pinnacle formation. × 37

FIGURE 3.—PORTION OF GRANITE BOULDER IN PINNACLE FORMATION
Skeletal crystals of magnetite-ilmenite are crossed by filaments of leucoxene. ×37

FIGURE 4.—CONTACT BETWEEN GRANITE BOULDER (LOWER HALF) AND MATRIX, PINNACLE FORMATION
Boulder chiefly of quartz, feldspar, and magnetite; matrix of same minerals with sericite and chlorite. ×37

FIGURE 5.—MASSIVE GRAYWACKE FACIES OF PINNACLE FORMATION
Q — quartz; F — feldspar; sericite, magnetite, and chlorite are also present. ×37

FIGURE 6.—SCHISTOSE FACIES OF PINNACLE FORMATION
Q — quartz; F — feldspar; sericite, magnetite, and chlorite are also present. ×37

brown-weathering dolomite in a matrix of gray-weathering white calcite. These patches range from the size of pin heads to several inches across. In many places the dolomitic and calciferous facies are interbedded; the latter occur as relatively small lenses. Small criss-crossing quartz veins are common in most outcrops; in some they are very abundant.

The dolomite is usually sandy; in some outcrops in the northeastern part of the Milton quadrangle, the sand is so abundant that it is more properly a sandstone with a dolomitic or siliceous cement. On the hill about half a mile south of School No. 13 in this same area, fresh reddish-brown crystals of orthoclase up to a quarter of an inch across are scattered through the dolomite. In places the dolomite is finely conglomeratic; the pebbles are not abundant. Commonly they are of quartz, but in some outcrops there are small pebbles of igneous rocks identical with some of those found in the boulder conglomerate of the Pinnacle formation. Some outcrops show numerous slate fragments. Locally there are many thin, slaty, siliceous, wavy interbeds which are more resistant to weathering than is the dolomite. Other evidences of stratification are rare. In a few outcrops tubular masses or seams of nearly pure chlorite 1–2 inches thick apparently parallel the bedding. Some exposures contain considerable hematite which is always concentrated at or near the upper contact. In one locality, the hematite is so abundant and in such large grains that superficially the rock resembles a conglomerate or a breccia. The hematite appears to be clastic, both in the hand specimen and in thin section. It may have been deposited as limonite which was subsequently dehydrated by metamorphism.

Flow structure is present in many outcrops; it is particularly noticeable where the limy facies is interbedded with the dolomitic facies. Even where the dolomitic facies alone is present, most of the beds have been so contorted by flowage and complex drag folding that the structure can be determined only by reference to the underlying and overlying beds. This complex drag folding is excellently displayed about half a mile north of South Franklin and in the northeastern part of the Milton Quadrangle.

One of the largest and best-exposed outcrops of the White Brook in Vermont is at the southern end of the large anticlinal inlier 3 miles northeast of the village of Fairfield. It caps a small prominent hill; on the northeastern side erosion has carved out an amphitheater-shaped lowland. In this amphitheater the underlying Pinnacle formation is exposed along the axis of the southward-plunging anticline. Outside the amphitheater, an irregular zone of pink, violet, and white marble can be traced around the nose of the fold. To the west the contact of the dolomite with the younger West Sutton slates is gradational. Another large expanse of dolomite, in the form of a dome in which the dolomite has not been breached, is found on Sheldon Hill, 3 miles to the northeast.

The White Brook dolomite in Vermont occurs as a homoclinal belt from the international boundary southward to the extreme northern part of the Mt. Mansfield quadrangle and as anticlinal inliers west of this belt. Synclinal outliers east of the belt are unknown. The east-west extent of the original area of deposition appears to have been small.

Clark (1936, p. 143) states that the minimum thickness of this formation in Quebec is 20 feet. In Vermont it is often less. On the west side of the prominent hill a mile north of the village of West Berkshire, a full thickness of about 10 feet of the dolomite is exposed near the base of the steep cliff on the west side. In places the formation is missing. Near the top of the hill 1½ miles S. 57° W. of the crossroads at the village of East Sheldon the gray phyllitic facies of the West Sutton formation lies directly in contact with the coarse extremely massive arkosic facies of the Pinnacle. The identification of these formations is attested by a hematiferous zone between them. A test pit had once been dug here. One piece of the hematiferous rock collected at this locality averaged about 40 per cent hematite; the remainder was largely sericite. Whether the absence of the White Brook in this and other areas is due to nondeposition or to squeezing out by tight folding is not entirely conjectural. Westward from the homoclinal belt of outcrop the formation thins, in places to only 1–5 feet of sandy dolomitic strata. In none of the western outliers does it appear to be over 10 or 15 feet thick, and it is more commonly

absent than present if the great dearth of outcrops is any criterion. The writer believes that the homoclinal belt of White Brook is absent south of the Enosburg Falls quadrangle. This thinning of the dolomite westward and southward creates a problem because it is an excellent horizon marker; without it the separation of the West Sutton and Pinnacle is impossible in many places. Therefore, the horizon of the White Brook is indicated on the map (Pl. 1) even where there is no evidence for its presence and where, in the opinion of the writer, it is actually absent. However, the symbol used differs from that used where the dolomite is known or where the writer is reasonably certain it is present under cover.

West Sutton Formation

The type locality of the West Sutton formation (Clark, 1934) is in the vicinity of the village of West Sutton in Quebec (Fig. 1). Clark (1936, p. 143–144) states that:

> "For the most part it is an impalpably fine mud rock, varying in color from shades of medium and black gray to dark blue-black. . . . One of the most constant characteristics is the presence of hematite. This may be disseminated throughout the rock as invisible particles allowing the rock to yield a dark brown or dark purplish-red streak, or it may occur as discrete masses, usually concentrated at the base of the bed near the contact with the dolomite. In some places this hematite is so pure and abundant that it has been worked as an iron ore."

There are few places in Vermont where this description is apt. One, along the road about a mile east of the village of Sheldon, is a well-cleaved mudstone rather than a true slate. The White Brook dolomite separates it from the typical coarse-grained graywacke facies of the Pinnacle formation. Hematite is concentrated at the base; a test pit was dug many years ago. Elsewhere in Vermont, the West Sutton is as variable as the Pinnacle. The average grain size is considerably smaller than in the Pinnacle, but the range is the same. Most commonly it is a sandy slate or phyllite which may vary only within relatively narrow limits over large areas. In other areas, sandy or quartzitic interbeds are common. Coarse-grained graywacke interbeds may be few and thin or they may be so common that they make up the bulk of the formation, as is especially true of the area around Fairfield Pond northward to the Missisquoi River. In a few places pebble and boulder conglomerates are present, and in one large area it is a greenish quartzose schist. Occasional dolomitic lenses occur, rarely over a few inches thick and a few tens of feet long. They may occur in discontinuous lenses along the same horizon for some hundreds of yards.

The West Sutton south of the Missisquoi River is so similar to the Pinnacle that the two can be differentiated only because of the intervening White Brook. North of the village of Fairfield, some of the anticlinal inliers have been breached by erosion so that the White Brook forms an elliptical ring around a core of slate that is mapped as Pinnacle on stratigraphic position, but no criteria could be developed to distinguish between it and the slate (West Sutton) overlying the White Brook outside the ring. The difficulty is increased by the fact that the hematiferous character of the West Sutton disappears southward.

At the international boundary the West Sutton outcrops in a belt not over half a mile wide that continues between the villages of East Franklin and West Berkshire. Along the road between these villages are excellent exposures. It is a quartzitic schist composed of alternating thin beds of greenish slaty material and gray-green quartzite. The latter beds are usually 1–2 inches thick. The slaty beds are well cleaved; cleavage parallels bedding, and both are much contorted in places. Near the international boundary the formation is less like Clark's description than almost anywhere else. Southward this facies changes to black, gray, and green slates and phyllites. South of the Missisquoi River the belt widens to about 4½ miles. The eastern portion of this belt is chiefly the typical slaty and phyllitic facies. The western portion from Fairfield Pond northward to the Missisquoi is largely medium-coarse-grained graywacke identical with the graywacke facies of the Pinnacle. This is mapped as West Sutton chiefly because it is in contact with the Gilman on the west. To map it as Pinnacle would involve inferring one or more faults. South of Fairfield Pond, the proportion of slate to graywacke increases, and the formation becomes more typical. In the pasture just east of French School, 2½ miles east of St. Albans village, there are

thick lenses of boulder conglomerate similar to those in the Pinnacle formation in the northeast corner of the Milton quadrangle except that no slate boulders are present and the matrix is of slate rather than of graywacke. There is no direct evidence of the position of these conglomeratic beds in the formation, but the writer believes they are near the base. Just north of the Lamoille River, the West Sutton is a fine-grained black phyllite. Far to the south, a little east of Colchester Pond, outcrops of slates and schists are mapped as West Sutton because of stratigraphic and structural relations. The summit of Brigham Hill is Gilman quartzite; the secondary summit just across a little ravine to the west is composed of an intensely crumpled greenish quartzitic and schistose rock containing much vein quartz, identified as West Sutton. It is separated from the Gilman to the east by a thrust fault which passes through the ravine.

In general the West Sutton rests directly on the White Brook, but contacts are rare. The contact appears to be gradational in a few places. In some areas the White Brook appears to be missing, and the West Sutton rests on the Pinnacle. Exposed contacts between the West Sutton and the Gilman are even rarer; those seen are abrupt. Clark (1936, p. 144) estimated the thickness of the formation to range between 40 and 250 feet in Quebec, 40 feet being more common. The thickness in Vermont, particularly south of the Missisquoi, probably averages at least 250 feet, and 400 feet may be a better estimate. No section could be measured; rarely can it even be estimated, largely because the underlying White Brook dolomite is nowhere exposed close to exposures of the overlying Gilman quartzite, and in part because of much repetition by folding.

Gilman Quartzite

The type section of the Gilman (Clark, 1934) is at Oak Hill in Quebec (Fig. 1). Clark states that in Quebec it is usually a quartz sandstone of such fine grain it should be called a shale. While this is essentially true north of the Missisquoi River in Vermont, southward it is most often coarse-grained enough to be called a siltstone or fine-grained sandstone, although in some areas it is argillaceous. It is far more uniform than the West Sutton and the Pinnacle formations. There is some variation across the strike, but the major variations are never repeated. There is also some variation parallel to the strike, but these variations are not abrupt nor are they repeated. This is a decided contrast with the Pinnacle and West Sutton formations where variations both along and across the strike are repeated in many areas.

Four facies of the Gilman may be recognized from bottom to top: (1) a coarse to very coarse-grained sandstone or quartzite, (2) a fine-grained argillaceous siltstone, (3) a mottled argillaceous quartzite, and (4) a whitish relatively pure massive quartzite. Never can all four be recognized in one traverse across the strike; none is continuous throughout the area from north to south. The lack of continuity is due to faulting, depth of erosion, and variations in the kind of material deposited in different areas. No definite boundaries can be drawn as changes are gradational, both laterally and vertically. Much careful work is necessary before ready recognition of all facies is possible, but this separation aids mapping the Fairfield Pond thrust.

South of the international boundary, most of the 4 mile wide eastern belt of Gilman is a very fine-grained, uniform argillaceous quartzite or siltstone. It is usually a dull, medium to dark gray. Weathering affects it very little except to whiten it. Bedding is never well displayed and most often is absent. Cleavage is often moderately well developed. Large grains of pinkish-tan calcite occur in small tension fractures in the more silty quartzitic and less cleaved beds. The fractures range from less than an inch to several inches long. None are completely filled with the calcite. Locally they may be abundant, as in the newer road cuts along the east-west road south of Shingle Hill (north of Lake Carmi). Elsewhere they are sparsely distributed throughout the whole phase. This phase forms all but the extreme western portion of the outcrop of Gilman north of the Missisquoi River, it outcrops all around Lake Carmi. In Quebec it forms the bulk of the Gilman quartzite as described by Clark. Between the Missisquoi River and Bellevue Hill, 2½ miles southeast of St. Albans, it forms much of the eastern parts of both belts of Gilman quartzite. South of Bellevue Hill it be-

comes relatively less important, possibly because it is partially cut out by a fault. It is missing between the Lamoille and Brigham Hill, probably because erosion has not yet cut deep enough to expose it. South of Brigham Hill it forms the largest part of the outcrop belt of the Gilman.

From Cushman Hill, 2½ miles north of the Lamoille River, to about half way between the French School and Fairfield Pond, the eastern part of the belt of outcrop is a coarse- to very coarse-grained sandstone or quartzite which is occasionally finely conglomeratic. It is light to medium gray and almost always well bedded; the beds are as thin as an inch, but 1 to 2 feet is much the most common. Cross-bedding is rare. It is well exposed along the top of the escarpment just east of St. Albans. This phase grades both laterally and upward into the siltstone facies.

The third phase of the Gilman is, for the most part, mottled argillaceous quartzite. Brownish to reddish-brown or gray to white sandy spots are separated by darker, finer sand or grit with some shaly material in which are interspersed a network of fine graphitic filaments. The sandy spots are elongated in the same direction and are irregularly arranged. This elongation may be mistaken for bedding, but in some places good bedding is at a considerable angle to it. It is probably related to the cleavage, which is well developed only in the more argillaceous phases. Good bedding is extremely rare. The lower part of this phase is the most argillaceous; the sandy spots are small, and the shaly material and carbonaceous filaments make up the bulk of the rock. This phase resembles the brown, less massive and more argillaceous lower Cheshire (Cady, 1945, p. 526). It is well exposed on Bridgeman and Minster hills near the international boundary. Southward it forms the western part of the western belt of Gilman to the vicinity of Greens Corners, 4½ miles northeast of St. Albans. It may be present under cover along the front of the escarpment from Greens Corners to Bellevue Hill. Southward, its separation from the upper phase is difficult. It is well exposed along the road on the north side of the Lamoille River. With the siltstone phase, it forms the "thumb" of Gilman, extending northward from the Lamoille along the Brigham Hill fault.

The uppermost phase is a white, light-gray, or medium-gray, relatively pure (in most cases), coarse- to medium-grained massive quartzite closely resembling the upper Cheshire (Cady, 1945, p. 526). Nowhere north of Greens Corners is it extensive enough to mention. It caps Aldis Hill, St. Albans Hill, and Arrowhead Mountain. South of the Lamoille River it includes a large part of the exposed portion of the main belt of the Gilman to Brigham Hill; in many places erosion has cut deep enough to expose the underlying mottled phase. The upper part grades into the Dunham. This gradation is best seen north of the road on the south slope of Bald Hill and three quarters of a mile west of Bowman's Corners. The transitional zone is a dark-gray dolomitic sandy slate. Where the overlying Dunham has been removed by erosion the slaty character of the transitional zone makes recognition of the outcrops as Gilman very difficult. Below this slaty transitional zone the Gilman is slightly dolomitic to a depth of some tens of feet. The dolomitic character of the quartzite is well displayed on the hill west of the village of Milton. The slaty character of the transitional zone is also seen low on the eastern slope of St. Albans Hill. In many places, particularly north of Greens Corners, this zone is either absent or so thin as not to be particularly noticeable.

The successive facies tend to coarsen upward from the siltstone facies and, in general, toward the west.

The thickness of the Gilman in Vermont has not been computed. It is certainly thicker than the 850 feet given by Cady (1945, p. 526) for the Cheshire in the Bristol-Starksboro region, but this thickness probably includes only the two upper phases of the Gilman. Clark (1936, p. 146) by various indirect means arrived at a figure of 3000 feet, a reasonable estimate for those areas where none of it has been cut out by faulting. Of the various phases the argillaceous siltstone is much the thickest north of the Missisquoi River.

Dunham Dolomite

The type locality of this formation (Clark, 1934) is at Oak Hill, Quebec; the name comes from the village of Dunham (Fig. 1), 4 miles southwest of Oak Hill. It is a light buff to

chocolate-brown weathering dolomite in almost all exposures. On fresh surfaces it is most commonly blue gray or steel gray, but there are many variations. The upper part in many places is light pink. In the east-west dolomite valley immediately north of Brigham Hill large exposures of reddish-purple dolomite grade eastward into gray types. Just east of the village of Franklin is an exposure of yellowish dolomite; a mile north of the international boundary this same belt is cream-colored, but between these two yellowish types are large outcrops of the steel-gray type. In the eastward re-entrant just south of the road leading from St. Albans to Fairfield Pond are two outcrops of gray-weathering thinly laminated limestone and one outcrop of glistening white dolomite in a "sea" of the dark-brown weathering type. The gray limestone occurs in patches, the largest about 100 yards long and 30–40 feet wide. The thickness appears not to be over 6 to 8 feet at the middle. It rests on gray dolomite. There are large solution cavities in it. The glistening white dolomite lies near by; it has a pearly sheen along shear planes and is very striking. This is the only area where these two types have been observed.

The Dunham is usually slightly sandy; well rounded sand grains are irregularly distributed through it in most outcrops. It is always finely crystalline. Stratification is absent in many exposures but is well developed in others because of thin siliceous and shaly interbeds that occasionally are reddish. Where these interbeds are present, the dolomite beds range from a few inches to several feet thick. In the lower part of the formation, undulating bedding planes give the rock a wavy and mottled appearance. Schuchert (1937, p. 1023) states that this character is due to the shallow sea in which the beds were deposited and to the rippled nature of the deposits.

A few outcrops of the Dunham have well-developed cleavage, as on the lower part of the steep hill north of the point where the St. Albans-Fairfield Pond road crosses the Oak Hill escarpment; it is nearly a slate. Cleavage is also well developed in some outcrops along the fault on the east side of Bridgeman Hill near the international boundary where it is approximately horizontal. Cleavage is well developed only in zones of intense thrusting, but it is not present in all such zones. Quartz veins, usually less than half an inch thick, crisscross the Dunham in many outcrops.

The Dunham outcrops almost continuously along the Oak Hill escarpment from the international boundary nearly to the Winooski River. Where it occurs in more than one belt, it is repeated by faulting or, less commonly, by folding. Its greatest development is along the eastern edge of the Champlain Lowland in the Oakland region south of St. Albans Hill and in the east-west valley immediately north of Brigham Hill. Very small outcrops appear on St. Albans and Aldis hills; it also forms a klippe on the eastern part of the summit of Cobble Hill. In the vicinity of Colchester Pond it disappears beneath the Oak Hill slate and the Brigham Hill thrust sheet. It reappears in a small area in the Winooski Valley northwest of the village of Williston where the front of the Hinesburg thrust sheet is eroded back 2 or 3 miles to the east, thus exposing a number of formations of the Oak Hill succession.

No section of the Dunham could be accurately measured. Where the outcrops are broadest it is thrown into a succession of folds. In places all or part of it is cut out by thrusting. Clark (1936, p. 147) states that it ranges from 40 to 150 feet in Quebec; the thickest section is at Dunham. Along the east-west road running along the north side of the valley immediately north of Brigham Hill (on the south slope of Bald Hill) the breadth of outcrop is several hundred feet. Amount of repetition by folding cannot be determined; an estimate of 300 feet was made, but it could easily be twice that, as a fault forms the eastern contact. Elsewhere estimates of 350, 100, and 500 feet were made. Certainly it is thinner in the St. Albans and Enosburg Falls quadrangles than it is in the Milton quadrangle. Cady (1945) gives greater thicknesses for the Dunham in the Rosenberg succession south of Hinesburg. Schuchert (1937, p. 1033) gives a thickness of 560 feet for the Mallett and Winooski dolomites (equivalent to the Dunham) 3 miles southeast of Swanton. The greater thicknesses in the Rosenberg succession may be due, in part at least, to erosion in post-Dunham time. The Mallett (upper) member (Cady, 1945, p. 528–529) of the Dunham

found in the Rosenberg succession is either absent or not sufficiently distinctive in the Oak Hill succession to be mapped. Its absence may be due to erosion in post-Dunham time or to gradation into only slightly sandy beds. Schuchert (1937, p. 1025, 1034) noted an erosional break between the Mallett dolomite, the upper member of the Dunham in the Rosenberg succession, and the Parker slate in the above-mentioned Swanton region; pebbles and boulders of dolomite are found in the lower Parker slate. Since an unknown portion of the lower part of the Dunham in the Rosenberg succession is cut out by thrusting (Champlain thrust), a comparison of the thicknesses in this succession from Malletts Bay northward with the thicknesses in the Oak Hill succession between Brigham Hill and St. Albans is impossible. The evidence indicates that the Dunham thins eastward. The apparent northward thinning may be a component of this easterly thinning, combined with post-Dunham erosion.

Parker Slate

In the Oak Hill succession, the type section of the Oak Hill slate (here designated as the Parker slate) (Clark, 1934) is at Oak Hill, Quebec. Clark (1936, p. 148) describes it as a dark-brownish-gray or black tough slate with the stratification planes shown by limonitic flecks, and more properly a siltstone than a mudstone. In Vermont the formation is variable. In places there are very numerous interbeds of sand from a fraction of an inch up to 2 or even 3 inches thick. Locally there are beds of quartzite and dolomitic quartzite several feet thick interpreted as interbeds. One of the most prominent of these is seen east of the cemetery at East Highgate. Both contacts with the slate have never been observed where these thick beds of quartzite appear, so one cannot tell whether the quartzite is an interbed or a quartzitic member of the overlying Rugg Brook formation. In some areas the Parker is dolomitic; in others there are interbeds of dolomite from a fraction of an inch to several inches thick and rarely more than a few tens of yards long. In a few places it is pyritiferous. In some outcrops, particularly near thrust zones, it has been so indurated that it resembles the Gilman. The sandy phase reveals the stratification, but in other phases it is obscure.

In places the contact with the Dunham is sharp; a few outcrops show what appears to be an intergradation but could be a pseudoconformability produced by the deposition of thin dolomitic lenses near the base of the Parker slate, such as those occasionally seen higher in the formation. Evidence for a disconformity between the Parker slate and the Dunham in the Rosenberg succession has already been pointed out in the discussion of the Dunham. At Oak Hill Clark (1936, p. 148) notes minor unconformities of considerable angular discordance in the Oak Hill, and patchy distribution of the slate. A disconformity evidently cuts down through the Oak Hill slate into the Dunham. Schuchert (1937, p. 1026) noted an erosional break in the Rosenberg succession at the end of Parker deposition that accounts for the absence of considerable of the Parker in that succession. The Parker in the Oak Hill succession is commonly present in Vermont except where it may have been cut out by faulting. Considerable variations in thicknesses, apparently due to erosion, were noted. The hiatus at the top of the Parker slate apparently decreases southward. The much thinner Dunham reported in Quebec as compared with Vermont may reflect, in part at least, the effect of erosion in post-Oak Hill (Parker) time.

Clark does not give the thickness of this slate in Quebec, but he gives the combined thicknesses of the Oak Hill, the Scottsmore, and the Sweetsburg (including the Vail slate member) as 300+ feet (1934, p. 10). The Parker slate in the Oak Hill succession in Vermont cannot be accurately measured because it has been complexly folded in an extremely complicated manner (Pl. 2, fig. 2), both on relatively large and extremely small scales. In places it has been overturned so that the fold axes are horizontal. It is probably never much thicker than 100 feet.

Rugg Brook Formation

The type section of the Rugg Brook formation (Howell, 1929) is in the Rosenberg succession near the mouth of Rugg Brook southwest

of St. Albans. Here it is a salmon-colored conglomerate containing blocks of dolomite in a paste of round-grained quartz cemented by dolomite. The type section of its correlative (the Scottsmore of the Oak Hill succession in Quebec) is 2 miles northeast of Oak Hill, Quebec (Clark, 1936, p. 137). Clark (1936, p. 148–149) describes it as a dolomitic quartzite and conglomerate. He states that the conglomerate exposures all differ so that to select any one for a type locality would be misleading. In Vermont the Rugg Brook is essentially a quartzitic dolomite and breccia. The fresh dolomite is commonly light gray, the weathered surfaces usually salmon-colored similar to those of the Dunham. The largest outcrop of this nonconglomeratic facies occurs along the Fairfax-St. Albans road east of St. Albans Hill.

The brecciated or conglomeratic facies outcrops only in a few areas. The boulders are chiefly gray to white limestone and are commonly slabby. Some are several feet across, but most range from a few inches to about a foot in length and may be 3 or 4 inches thick. The best outcrop of this boulder conglomerate is on a knoll on the northeastern slope of Bridgeman Hill (Pl. 2, fig. 3). About a mile north of this knoll and in the line of strike is another type of conglomerate consisting of dolomite boulders in a sandy matrix.

The source of the gray and white limestone boulders is difficult to establish. According to Clark (1936, p. 149) there is little doubt that most of the boulders (in one outcrop) are derived from the Dunham. This, however, does not account for the limestone boulders. He reports only one occurrence of gray limestone (1936, p. 147) and one of white or yellowish dolomite in the Dunham. The writer has found only one area of gray and white variations of the Dunham and only two areas where creamy-white or yellowish variations occur. None of them closely resembles the limestone boulders in the breccia. No dolomites or limestones known in northeastern United States or southeastern Canada are older than the Dunham except the White Brook and its possible equivalent in west-central Vermont, the Forestdale marble. Certain calcareous varieties of the White Brook strongly resemble some of the boulders in the breccia, but the White Brook is separated from the Dunham by several thousands of feet of clastics, and there is no evidence for the great depth of erosion necessary to derive boulders from it. The angularity of the boulders suggests local origin.

Two sources are suggested: (1) The upper part of the Dunham, as originally deposited, may have contained much more gray and white limestone than present outcrops indicate. There is some evidence of an erosion interval in post-Dunham and pre-Rugg Brook time. Perhaps this part was removed by erosion, but it hardly plausible that a top member would be evenly stripped. (2) Schuchert (1937, p. 1025) reports bioherms in the Parker of the Rosenberg succession. None has been observed in the Parker of the Oak Hill succession, but only very small portions of the slate are exposed. Possibly the bioherms developed only in the upper part of the Parker and were removed by erosion. Schuchert's (1937, p. 1048–1050) explanation for the origin of the Upper Cambrian Rockledge breccia of the Rosenberg succession might also explain the origin of the Rugg Brook breccia.

The occurrence of the Rugg Brook as narrow discontinuous belts, sometimes in tightly folded synclines, makes mapping of the Parker and the Skeels Corners slates difficult in many areas because no lithologic criteria differentiate the slates. Moreover, the Rugg Brook is variable and resembles interbeds in the Parker and the Dunham. Some of the outcrops mapped as Rugg Brook may be relatively thick lenses of quartzite or dolomitic quartzite in the Parker. However, the change from slate to quartzite is abrupt at both contacts; lack of bedding in the quartzite, and commonly in the Parker, prevents determination of the structure in these outcrops. Where the brecciated facies is absent, the Rugg Brook is commonly indistinguishable from the Dunham. It lacks the wavy mottling, the siliceous interbeds, and many of the color variations of the Dunham, but these characters are not present in all phases of the Dunham. Structural criteria are commonly absent. Therefore, rocks identified as Rugg Brook in a few places, notably 1 or 2 miles northeast of Aldis Hill, 2–3 miles south of St. Albans Hill, and on the northwest slope of Brigham Hill, may be sandy phases in the upper part of the Dun-

ham and may represent an eastern facies of the Mallett member (Cady, 1945, p. 528–529).

The limestone boulder facies of the Rugg Brook looks like the Rockledge conglomerate above the Skeels Corners slate and below the Hungerford slate (Billings, 1948, personal communication). The Parker, Skeels Corners, and the Hungerford slates are similar lithologically, all are intricately folded (Pl. 2, fig. 2) and in some areas are distinguishable only by the presence of the Rugg Brook or the Rockledge formations between them, both of which may occur as homoclines or as tightly folded synclines in the Parker and Skeels Corners slates respectively. If it were not for critical exposures in two areas, on the northeast slope of Bridgeman Hill, and in the Winooski valley about a mile northwest of Williston, the conglomerate might everywhere be mapped as Rockledge. Correlation of the breccia in these two areas, which are 40 miles apart, with the Rockledge would assume that the Rugg Brook and the Skeels Corners are absent in both regions but present in much of the region between them. Furthermore, such a correlation would not account for the dolomite boulder conglomerate outcrops northeast of Bridgeman Hill and in Quebec (Clark, 1936, p. 148–149). Dolomite boulders are not known in the Rockledge of the Rosenberg succession. Detailed studies of the eastern belt of the Rosenberg slice in the St. Albans area by Alan Shaw (1948, personal communication) indicate the presence of formations comparable to and the stratigraphic equivalents of the Parker, Rugg, Brook, and Skeels Corners formations west of and parallel to the Oak Hill escarpment. Moreover, the formations of the two successions occur in apparent unbroken stratigraphic order north of Green's Corners (4 miles northeast of St. Albans). Such sequences cast doubt on the validity of the Oak Hill thrust as described by Clark (1934, p. 4, Fig. 2) as well as on the identity of the Oak Hill succession as distinct from the Rosenberg succession. However, other thrusts, to be discussed later, may, in part at least, take the place of the Oak Hill thrust in Vermont and may be related to it in time. The apparent structural concordance of the Rosenberg and Oak Hill successions does not necessarily deny the presence of a thrust between them.

Skeels Corners Slate

The type section of the Skeels Corners slate (Howell, 1939) is in the Rosenberg slice near Skeels Corners 7¼ miles north of St. Albans. The type section of its probable correlative (the Sweetsburg slate) in the Oak Hill succession of Quebec is at Oak Hill (Clark, 1934). Clark's description (1936, p. 149–150) of the Sweetsburg is not very applicable to the Skeels Corners. The latter is so much like the Parker that, where fossils are lacking, they cannot be distinguished. In the Rosenberg slice, the Skeels Corners contains a peculiar hornlike fossil, *Bovicornellum vermontense* Howell, but this fossil has been found in only one area in the Oak Hill succession, along the Fairfax-St. Albans road directly east of St. Albans Hill where the Skeels Corners forms the west slope of the low northward-trending ridge. The Parker slate forms the east slope of this ridge; the Rugg Brook outcropping on the top separates the two. The relative ages of the two slates with respect to the Rugg Brook are clearly determined from cleavage-bedding relationships. This is the only area of Skeels Corners definitely recognized by the writer in the Oak Hill succession in Vermont. The thickness of the Skeels Corners slate has not been determined. The Skeels Corners, and the Parker, are indistinguishable from various slates of the Rosenberg succession beneath the Champlain Lowland west of the Oak Hill escarpment. Clark (1934, p. 4, Fig. 2) reports a thrust fault, the Oak Hill thrust, west of the Sweetsburg slate (1936, p. 141, Fig. 3) in Quebec, but such a thrust, is not demonstrable south of the international boundary. Thus a considerable thickness of the Sweetsburg or its correlative, the Skeels Corners, may underlie parts of the area immediately west of the Oak Hill escarpment.

Undifferentiated Rocks of the Eastern Terrane

The writer studied the formations in the northwestern part of the Jay Peak quadrangle, the southeastern part of the Enosburg Falls quadrangle, the western part of the Mt. Mansfield quadrangle, and the extreme eastern part of the Milton quadrangle. The White

Brook dolomite could not be traced south of the Enosburg Falls quadrangle, and consequently the West Sutton and the Pinnacle formations could not be differentiated in that area. The Tibbit Hill schist appears in small anticlinal inliers in the Pinnacle as far south as the Lamoille River. The Pinnacle terrane seems to pass imperceptibly eastward into the more metamorphosed rocks that flank the Green Mountains. Therefore the present report was limited eastward to the area in which the various formations of the Oak Hill succession could be identified over sizable areas. Nevertheless considerable information was gained from the reconnaissance study of the eastern terranes.

The moderately coarse-grained Pinnacle graywacke occurs in numerous synclinal outliers in the Tibbit Hill schist north of the Missisquoi River. In the northwestern part of the Jay Peak quadrangle, the Tibbit Hill is bordered on the east by very fine-grained greenish chloritic sandstones superficially like the Tibbit Hill in many outcrops. This sandstone terrane passes eastward into fine- to coarse-grained chloritic, micaceous, quartzitic, and sometimes graphitic schists which make up the bulk of the Green Mountains in this latitude. These schists are coarser-grained and more quartzitic in the heart of the Green Mountains than they are either to the east or the west.

South of the Missisquoi a short distance, the coarse-grained facies of the Pinnacle grades southward in a short distance into fine-grained graywacke and greenish to blackish-gray slates and phyllites. In places numerous lenses of dolomite up to a foot thick and a few yards long pinch in and out. The main belt of Tibbit Hill borders the Pinnacle on the east for a few miles south of the Missisquoi and then plunges beneath the Pinnacle. A traverse along the road from Fairfield to East Fairfield reveals a transition from the blackish-gray or greenish-gray slates a little below the White Brook dolomite to silvery micaceous schists even before Black Creek is reached. Locally the schist contains quartz lenses and stringers, well exposed in many new road cuts from Fairfield Station to East Fairfield, where typical beds of the greenish-gray graywacke facies are exposed along the stream and on the hills to the north and south. On the hill south of East Fairfield the Tibbit Hill is exposed in anticlinal inliers. Eastward toward Bakersfield the graywacke beds become fewer and less typical and merge with the schist terrane. The fine-grained chloritic sandstones east of the main belt of Tibbit Hill farther north are not encountered in this traverse.

The fine-grained slaty to schistose facies of the Pinnacle continues to about the northern border of the Mt. Mansfield quadrangle where they gradually become coarser-grained and pass into the more typical graywacke facies. This latter facies is particularly prominent in the Fairfax, Fairfax Falls, and Westford regions, where it forms ridges markedly sculptured by glacial action. In many places it is sparingly conglomeratic; the pebbles are chiefly quartz grains the size of peas. To the west, nearer the Brigham Hill fault, the rocks are considerably finer-grained, chiefly black and green slates but with many interbeds of graywacke in some places. These slates are believed to belong to the West Sutton formation.

Eastward along the Lamoille River from Fairfax to Jeffersonville the same type of change is noted. Beds of graywacke become fewer, although typical graywacke can be recognized for some distance east of Cambridge. Interbeds of finer-grained material become more numerous and increasingly metamorphosed; quartz-mica schists are common, and graphitic phyllites and quartzites occur. The number of quartz lenses and stringers increase markedly.

South of Westford and Essex Center, typical graywacke beds of the Pinnacle can be easily recognized as far as the Winooski River; the finer-grained facies again becomes more prominent in this direction. The Pinnacle is represented south of the Winooski as far as the latitude of Rutland; it pinches in and out along the axis of an anticline in an hour-glass pattern, although the structural complexities make it a little harder to trace (Cady, 1948, personal communication).

One unsolved problem is the relationship of the rocks that form the Green Mountains to the rocks of the Oak Hill succession as defined by Clark (1934, Fig. 2, p. 4, 15) who identified the Brome thrust as the eastern border of

the Oak Hill slice. He placed this thrust, which he defined as an underthrust, at the eastern limit of the main outcrop belt of Tibbit Hill schist, which, in Quebec, everywhere intervenes between the main outcrop belt of the Pinnacle formation and the Sutton schists (the schists which form the Green (Sutton) Mountains in Quebec). He also stated (1934, p. 10–11),

"... the writer has come to the conclusion that the schists (Sutton) represent not only the Oak Hill series, but also a thick body of sediments younger than anything in that series, including not only other Lower Cambrian beds, but Upper Cambrian, and possibly Ordovician, as well. ... Here and there throughout the western part of the Sutton schist area similar rocks occur, which fact indicates that in anticlinal folds the underlying Oak Hill beds are brought up to the surface. Nowhere east of the main Sutton Mountains anticlinal axis do we find such indications."

Hence, he believes that the lowermost strata in the Sutton (Green) Mountains are correlatives of the Oak Hill series, that the Tibbit Hill schist is the oldest formation in southern Quebec, and that most of the rocks in the mountains are post lower Cambrian.

In Vermont the eastern border of the main belt of outcrop of the Tibbit Hill may be traced from near the international boundary post No. 603A across the northwestern part of the Jay Peak quadrangle to an indefinite point about 2 or 3 miles south of the village of Enosburg Falls, where it plunges beneath the Pinnacle formation. The contact between the Tibbit Hill and the Green Mountain rocks is always vertical or nearly so, and the bedding parallels the foliation. It is, therefore, impossible to determine the relative ages from the structural relations. The trace of the contact is like the traces of the contacts of the lower formations of the Oak Hill succession. Southward from where the easternmost part of the main outcrop belt of the Tibbit Hill plunges beneath the Pinnacle, the latter merges with the schist terrane of the Green Mountains, but a sharp contact is not recognizable because the fine-grained chloritic sandstones which form the westernmost outcrops of the latter terrane farther north grade laterally southward into schists which are indistinguishable from the fine-grained facies of the Pinnacle which is so prevalent in this area. Traverses across the southward projection of the Tibbit Hill-Green Mountain rocks contact indicate that the Green Mountain schist terrane passes into that of the Pinnacle formation without any break. There is no evidence here for the Brome thrust. These Green Mountain rocks appear younger than the Tibbit Hill schist and, if so, are the equivalents of part or all of the Oak Hill succession. On the steep eastern face of Peaked Mountain about 3 miles east of Bakersfield a zone of coarse-grained marble may be the equivalent of carbonate rocks of the Oak Hill succession. The evidence in Vermont thus supports the main points in Clark's conclusions, but the tectonic limits of the slice, if any, are not to be found in the western flanks of the Green Mountains.

Cady (1945, Fig. 5, p. 563) shows the Taconic thrust passing west of the Green Mountains and east of the Hinesburg and Oak Hill slices, but there is no local evidence for it in the area of this report. The present consensus of geologists who have studied the problem is that the Taconic thrust extends far beyond the Taconic Range. Various geologists, among them Kay (1937, p. 216, Pl. 5) and Cady (1945, p. 569; Fig. 6, p. 568), have suggested that the Taconic thrust connects with the thrust that has been traced southwestward from the Quebec City region into the Granby slice (Clark, 1934, Fig. 2, p. 4, 14) southeast of the St. Lawrence River.

Interpretation of the Stratigraphic Record

The geologic history of this part of Vermont began probably in earliest Lower Cambrian time with the extrusion of the basic lavas that comprise the Tibbit Hill schist. At the end of the volcanic activity the surface had little relief and was covered by the strata of the Pinnacle formation. Sedimentation generally began with coarse-grained argillaceous sands. Some weathering and erosion may have preceded the deposition of the sands, for the Tibbit Hill is not vesicular at its contact with the Pinnacle in any place observed. Any regolith must have been carried to another region, probably to the east.

The parent rocks of the igneous boulder facies and the graywacke facies of the Pin-

nacle and West Sutton formations were the same, for the mineralogy of all three is similar. The source region was one of little diversity, chiefly granitic and syenitic rocks. The angularity of the grains, their relatively small reduction in size compared to the corresponding minerals in the boulders, and the incomplete sorting suggest that the site of deposition was not far from the place of origin. The source is believed to have been west or northwest of the original area of deposition and now lies beneath westerly overlapping higher strata and perhaps one or more of the thrust slices of northwestern Vermont and southeastern Quebec. It apparently lay west of the original geosynclinal area.

A western or northwestern source is indicated because (1) many of the western anticlinal inliers of Pinnacle are conglomeratic, but none of the eastern ones. (2) North of the Missisquoi River the main belt of Pinnacle is considerably coarser than the more schistose facies in the eastern synclinal outliers in the northwestern part of the Jah Peak quadrangle. (3) West of the homoclinal belt of the White Brook dolomite which extends the full length of the Enosburg Falls quadrangle, the western anticlinal inliers are coarser-grained than the eastern inliers, and the latter are coarser-grained than the main belt south of the line of facies change in the latitude of West Enosburg (Pl. 1). (4) The main belt of Pinnacle north of the above line of facies change is generally considerably coarser than that to the south; the predominantly argillaceous facies is absent both in northern Vermont and in Quebec. An exception is the eastward-jutting salient of coarse and, in many places, finely conglomeratic Pinnacle a little southwest of the village of Fairfield. However, the most conglomeratic facies of the Pinnacle is in the anticlinal inliers west of this salient, and eastward the Pinnacle quickly grades into finer-grained clastics east of Fairfield River. At Soule School, it is a typical phyllite. (5) The White Brook dolomite is extremely sandy locally and is conglomeratic in the westernmost anticlinal inliers, whereas it is nowhere so in the homoclinal belt. (6) The distribution of the coarse-grained facies of the West Sutton formation with respect to the fine-grained facies corresponds to that of the Pinnacle in most respects. The one important difference is that the West Sutton is coarser between the latitude of St. Albans and the Missisquoi River than it is farther north.

The upland area from which the sediments of the Pinnacle formation came must have been sufficiently high during most of the time involved for disintegration to exceed decomposition. Lack of vegetation in Cambrian times would tend further to make disintegration predominant over decomposition, and would hasten the removal of freshly disintegrated rock. The gradients were steep enough for the streams to carry coarse sands and, at times, pebbles and boulders up to 15 inches or more across.

The distribution of the coarse and fine facies of the Pinnacle with respect to the western shore line of the geosyncline is not consistent, but most, or all, the discrepancies may be explained in one or more of the following ways: (1) Since the shore line was gradually progressing westward during Pinnacle time, the lowest Pinnacle in the east may be older than the lowest Pinnacle in the west. (2) Some of the apparent variations across the strike are structural repetitions of single layers of fine sediment deposited on coarse detritus. (3) Some of the variations both along and across the strike are due to differences in time of attainment of a profile of equilibrium for the various grain sizes. These processes, combined with the intermittent sinking of the geosyncline which destroyed the successive surfaces of equilibrium produced diastems in the stratigraphic record and abrupt changes in the type of material deposited.

The writer has elsewhere implied that the Pinnacle is a marine-deposited sediment. The boulder conglomerates give no clue of their origin. No reliable criteria have been developed to distinguish marine from fluviatile boulder conglomerates. The formation, except quite locally, lacks cross-bedding and presumably was largely deposited below effective wave base. Furthermore, it is too well sorted, despite its variability, to be classed as a deposit made by streams alone in a piedmont environment. Distribution of the textural facies parallels the west shore of the geosyncline. Moreover, the succeeding dolomite is undoubtedly marine. The Pinnacle and the later formations of the succession seem to be marine formations deposited in the neritic zone. Because of intermittent

sinking of the floor of the geosyncline thick deposits accumulated.

During earliest Pinnacle time, the shore line was probably youthful. The continued rapid deposition of the clastics shallowed the waters; the sea floor was sinking and the sea advancing westward, but not fast enough to prevent surfaces of equilibrium from being established toward the close of the period. Where equilibrium was first established, permanent deposition ceased for a time. Elsewhere deposition continued, and fine clastics were laid down.

Eventually clastic deposition ceased, and conditions favored deposition of the White Brook dolomite. Occasionally a little mud was washed in to become interbedded; small pebbles and sand were mixed with the dolomite in the near-shore areas. In a few places, possibly at deltas opposite the mouths of streams, the dolomite was not deposited at all. This might explain the absence of the White Brook in some of the western anticlinal inliers and locally in the homoclinal belt of outcrop.

A slight uplift of the land area to the west caused cessation of dolomite deposition and renewed deposition of clastics which make up the West Sutton formation. The change appears to have been rapid because there is little intergradation between the two formations. Iron oxide in considerable abundance accumulated in the northern part of the area in late White Brook and early West Sutton times. The iron oxide is now hematite but probably was deposited as iron hydroxide which was dehydrated during the metamorphism of the rocks. The events of West Sutton times differed little from those of Pinnacle time. Decomposition at the source seems to have been more effective in early West Sutton times, but this effectiveness decreased as time advanced, for the formation changes from fine- to coarse-grained upward and westward.

There is an abrupt change from the medium-coarse West Sutton graywacke to the finer Gilman quartzitic siltstone between Fairfield Pond and the Missisquoi River. In the latitude of St. Albans the change is from a blackish to greenish sandy mudstone to a medium-coarse-grained and relatively pure quartz sand; and a little farther south from a gray-black mudstone to coarse sands. Wherever observed, the change is abrupt, though there is no evidence of any significant erosion interval.

This change from West Sutton types of sediments to Gilman types may be explained by assuming a general uplift which caused the strand line to move easterly. The previously deposited strata were reworked by wave and current action in the littoral and upper neritic zones, and the sedimentary particles became sorted both as to size and constituent materials. The coarser (more resistant) quartz grains were deposited relatively close to shore, to form the Gilman quartzite; the finer (less resistant) grains, including feldspar, were transported eastward and deposited at somewhat greater depths in the neritic zone, apparently east of the area of outcrop of the typical Gilman. Such eastern facies deposited in Gilman time are apparently now eroded from the terrane of undifferentiated rocks immediately east of the area mapped. This distribution of facies is inferred from the distribution of reworked sands found higher in the stratigraphic section in the Rosenberg succession (Cady, 1945, p. 534). Because the Gilman sands represent the initial separation of quartz sands from the finer material, they are believed to have graded laterally eastward into the fines. Subsequent sandstones, such as those higher in the Rosenberg succession, were formed merely by reworking of the older reworked sands; thus they grade laterally eastward into dolomites and limestones rather han into finer clastics.

As the relief at the strand and west was reduced and approached base level, the supply of reworked sand, silt, and mud diminished and was finally cut off, and deposition of dolomite began again. This time the transition was not abrupt; the contact between the Gilman and the Dunham is gradational. The tranquil conditions which favored dolomite deposition continued far longer than they had in White Brook time. Occasionally small amounts of mud and silt were washed in to become interbedded. The wind also transported coarse rounded glassy grains of sand and dropped them in the sea (Schuchert, 1937, p. 1024). Toward the close of Dunham time, there is evidence of emergence of the geosyncline and formation of elevated tracts to the east within its belt, sufficiently to allow the upper and eastern portions of the dolomite to be removed

by erosion, especially in the northern part of the area.

With resumption of geosynclinal subsidence, the sea returned, and the regolith, which had developed probably in the elevated eastern tracts (Cady, 1945, p. 533, 534, 538–539) on earlier argillaceous strata, was washed in. These were the muds which now form the Parker slate.[2] Again the sea retreated, for an unknown interval, and returned at least once. The evidence for the length of the latter interval is missing from the Oak Hill succession in Vermont, but in Quebec and in the Rosenberg succession to the west information suggests that at least the remainder of early Cambrian and possibly medial and late Cambrian and early Ordovician times may have been involved.

Thus at least four cycles of erosion and related sedimentation produced sediments as follows: (1) local gravels, feldspathic sands, silts, and muds followed by dolomite (Pinnacle and White Brook formations), (2) muds, silts, and sands (West Sutton formation), (3) sands followed by dolomite (Gilman and Dunham formations), (4) mud followed by sand and dolomite (Parker and Rugg Brook formations).

Reworked sediments first appeared at the beginning of the third cycle and formed practically all of the clastic sediments thereafter; apparently the western granitic-syenitic terrane was covered with earlier strata. The reworked sands were derived chiefly from the western arenaceous facies, and the reworked silts and muds came from the eastern argillaceous facies of the initial (unreworked) deposits. The carbonate facies were formed out of range of the spread of the clastics; hence during the time of easterly spread of the initial clastics they formed in more eastern areas, but later, when clastics came from both directions, optimum conditions for their formation prevailed in central or western belts, approaching the more stable western margin of the geosyncline. Thus in the typical cycle they commonly follow the deposition of sands rather than silts and muds.

[2] These muds could not have been derived from reworked sands along the strand line. Present distribution also precludes this as the Monkton pinches out to east into dolomites (Cady, 1945, p. 531) and the Parker pinches out to west into dolomites at same stratigraphic level. In other words the shales are "boxed" into an eastern source much as the sands are "boxed" into a western source.

Intrusives

Intrusives in the Oak Hill succession are limited to a few very short thin dikes and irregular stringers, chiefly of granite pegmatite. The pegmatite dikes, never conspicuous, occur chiefly in the Tibbit Hill and the Pinnacle formations. Two deeply weathered dikes, only a few inches thick, possibly aplites, were noted in the Enosburg Falls area. One dike, about 2 feet thick, occurs in the Parker slate northwest of Brigham Hill. It is a moderately fine-grained porphyritic basaltic rock with relatively large phenocrysts of hornblende. None of these dikes could be traced more than a few feet.

Irregular knots, stringers, and veins of quartz are extremely common in the eastern rocks. Much of this quartz may be due to the effects of regional metamorphism, but some, especially that in crosscutting veins, was apparently derived from intrusives at greater depth. In the dolomites, introduced quartz appears chiefly as veins which commonly strike at right angles to the strike of the bedding, particularly on overturned limbs.

Metamorphism

All the rocks of the Oak Hill succession show some degree of metamorphism. Degree of metamorphism increases eastward from very low grade on the west to low-middle grade in conformity with the development of middle-high grade metamorphics high along the axis to the Green Mountains.

The amount of quartz in bands parallel to the foliation increases easterly. This widespread occurrence of foliate structures in conjunction with the general absence of intrusive igneous rocks favors regional (dynamothermal) metamorphism rather than thermal metamorphism. Chlorite or carbonate occurs with the quartz, particularly in the easterly formations. In many places in the fine-grained facies of the Pinnacle and West Sutton formations, the apparently segregated bands of quartz are so numerous that incompetent slates have been transformed into highly resistant quartzose schists. One such locality is on the secondary summit west of

the main summit of Brigham Hill; another is on the hill (elevation 1700 feet) south-southeast of Sheldon Hill in the middle of the Enosburg Falls quadrangle. Some of the quartz bands have been folded and sheared. Other bands in the same outcrops lack these features and may have been formed during the late stages of deformation, or possibly they represent intrusive quartz veins formed after the release of stresses.

Structural Features

General Setting

The major structural features of northwestern Vermont are shown in Figure 2 (modified from Cady, 1945, Fig. 5, p. 563). The fold axes in western Vermont south of the St. Albans area strike about north-south, but in places as far north as Bellevue Hill (2¼ miles southeast of St. Albans) the trend may be as much as N. 10°W. The change in the trend north of Bellevue Hill to approximately N.30°E. is accomplished in about 5 miles. Thus the axis of a regional salient of the Appalachian mountain system crosses the major structures of this area so that they lie about equally on the two flanks of the salient. A small anticlinorium, a great homocline, and four thrust faults constitute the major structures. North of the axis of the regional salient, the anticlinorium passes into a homocline that may be considered the western half of a greater anticlinorium whose axis lies largely east of the area mapped; to the south, the smaller anticlinorium passes beneath thrust slices apparently formed within the western limb of the greater anticlinorium. Thrust faults lie both east and west of the smaller anticlinorium; the Brigham Hill thrust to the east is developed largely south of the axis of the regional salient, and the Fairfield Pond thrust to the west, which extends northerly into the western part of the great homocline, is formed largely north of the axis of the regional salient. Several klippen or similar structures that lie west of the Oak Hill escarpment are believed to be genetically related to the thrusts.

Anticlinorium

The oldest beds exposed in the smaller anticlinorium lie roughly athwart the axis of the regional salient. To the north and to the south of the latter axis, the strata of the anticlinorium become younger; thus the axis of the anticlinorium plunges northerly to the north of the salient and southerly to the south. The eastern limb of the anticlinorium merges with a subdued and poorly defined synclinorial tract that extends a short distance north of the axis of the regional salient for several miles; south of the salient axis, the anticlinorium is bounded on the east by the Brigham Hill thrust. The anticlinorium is overlain by a western extension of the Brigham Hill thrust slice from Brigham Hill south across the Winooski River valley. The western limits of the anticlinorium are roughly coincident with the Oak Hill escarpment. West of the escarpment, the younger formations of the succession occur in an apparently broad syncline that is probably nearly complete in the Milton region but may have been partially eroded elsewhere. Many minor folds occur in this syncline west of the Oak Hill escarpment; they are well displayed in the Milton region northward to Bellevue Hill. Dips in the Dunham range from 40° to 80°, but the variation may be as much as 30° in a few feet. The anticlinorium is overturned to the west. The dips on the east limb range from 40° to near vertical; dips higher than 70° are common. On the west limb along the Oak Hill escarpment they range between 60° and vertical; locally they are westerly.

Homocline

The younger beds are to the west, continuous northerly with those on the west limb of the anticlinorium, but there is no suggestion of a major repetition of younger formations to the east in the northern part of the area, despite the strong folding which produced a predominance of steep dips and considerable overturning. This, plus the fact that the formations beneath the Gilman are not especially thick and yet outcrop in broad belts in a region of considerable topographic relief, indicates that no major folds such as the anticlinorium to the south were formed. A subsidiary anticline in the western part of the homocline is truncated by the Fairfield Pond thrust. The rocks to the east were crumpled into a series of relatively small doubly plunging closed folds whose axial planes in general dip steeply east, although

dips as low as 30° E. were noted. The numerous small anticlinal inliers and synclinal outliers, and the superabundance of large drag folds that may best be seen in the White Brook dolomite, are repeated many times along a line across the strike and indicate the above type of structure. Half a mile northeast of South Franklin, at least nine narrow belts of White Brook can be crossed in a nearly east-west traverse of about 200 yards, and 15 bands can be crossed in about 400 yards if a winding traverse is made. The gentle westerly dip of this entire folded belt is reflected in that the formations become successively younger to the west without a corresponding decrease in elevation. This general westerly dip, combined with the general southward plunge of the homocline and of the steeper southern nose of the anticlinorium, has allowed erosion to cut stratigraphically deeper in the east and the north than in the west and the south.

Steep eastern dips are general in the homoclinal area except in the north along the front of the Oak Hill escarpment. Westerly dips from 25° to 60° and occasionally steeper are common in the Bridgeman-Minster Hill area. Farther east, and to a lesser extent to the south, the dips steepen and are almost always easterly; rarely are they less than 50°; more often they are near the vertical. Overturning to the west is common; in the White Brook dolomite, the dips on the overturned limbs may be less than 30°E. The complexity of the folding is observed best in the White Brook formation. Relatively flat or moderate southerly or low northerly dips in some places are found along or near the axes of folds. The regional strike in the homoclinal area is more easterly than in the anticlinorium. While the average is between N. 25°E. and N. 30°E., locally the strikes vary considerably. The type of structure found in the homoclinal area continues southerly east of the trace of the Brigham Hill thrust in the undifferentiated rocks east of the area mapped in detail.

The transition from the anticlinorium in the south to the homocline in the north begins at about the same latitude that the Brigham Hill thrust dies out, and the homocline probably does not completely replace the anticlinorium until the point is reached where the thrust actually does die out. The thrust seems to follow a zone of strong flexure east of the anticlinorium that becomes imperceptible in the homocline.

Thrust Faults

General statement.—Two major thrusts, the Brigham Hill and the Fairfield Pond, are clearly recognizable in the area mapped north of the Winooski River, and one, the Hinesburg thrust, is recognizable only south of the Winooski. A fourth thrust, here termed the Arrowhead thrust, may not be a separate thrust; it may be a continuation of the Fairfield Pond thrust or it may be related to the Brigham Hill thrust. The Oak Hill thrust (Clark, 1934, Fig. 2, p. 4) cannot be recognized with certainty in Vermont. The possible northward extension of the Fairfield Pond thrust crosses the international boundary a mile or so too far east (and is in the Gilman formation rather than at the base of the Oak Hill escarpment) to be the southward continuation of the Oak Hill thrust as mapped by Clark.

Brigham Hill thrust.—The Brigham Hill thrust slice overlies the southern extremity of the anticlinorium from Brigham Hill south, and the trace of this thrust extends northerly along the east limb of the anticlinorium and disappears in the homoclinal succession. Klippen west of the Oak Hill escarpment and of the anticlinorium may be erosion remnants of the thrust slice. Northward from Brigham Hill along the eastern edge of the anticlinorium the fault is apparent because of the omission of strata. Here the Dunham is in contact with the West Sutton slates east of the thrust; the Gilman is missing. Valleys are developed continuously along the trace of the fault northward from the east-west valley just north of Brigham Hill. The dolomite extends down to, or very near, the foot of the western slopes of the valleys, slate or graywacke down to or near the foot of the eastern slopes. Despite the lack of exposure of the thrust plane along these valleys, the fault dips to the east at a moderately high angle, possibly 20°, because the eastern slope of Bald Hill is steep; the thrust plane cannot dip more gently than this slope. Northward the thrust swings northwesterly and approaches the axis of the anticlinorium so that it first cuts out the Dunham, then the

Gilman, and, finally, in part, the West Sutton formation. It dies out in the latter formation somewhere to the north.

Southward the fault is exposed on the west side of Brigham Hill. It cannot be traced on the eastern slope of the hill, but its presence is inferred because of the rapid narrowing of the Gilman outcrop between the Dunham on the west and the West Sutton to the southeast. A small ravine separates the summit of Brigham Hill from the slightly lower secondary summit to the west. In the northern end of this ravine, the greenish quartzose schist (here mapped as West Sutton) which forms the secondary summit rests upon the Dunham dolomite. The thrust plane is nearly flat. Gilman quartzite caps the main summit of the hill; this cap is isolated from the Gilman to the east and is surrounded by the Dunham on the north, east, and south. It is separated from the West Sutton on the west by the thrust that passes through the ravine. There are two possible explanations for this isolated cap of Gilman. It might be an inlier in the Dunham dolomite; it would thus be in place and below the thrust plane. It might be a fault outlier—a klippe—a remnant of the mass of Gilman that was thrust over the Dunham here. The writer favors the latter interpretation because the thrust plane where observed in the ravine dips too gently to carry over the Gilman to the east, the dips of the Gilman on its eastern and western edges are too low to bring it under the Dunham in the distance available, and in one place, although the contact cannot be observed, the Gilman apparently overlies the Dunham. The Dunham is believed to be in place below the thrust plane because rocks identified as West Sutton lie above it, and the outcrops of Dunham on the hill tie in with the western belt of dolomite which extends down from the north. The latter cannot be traced all the way through because no outcrops appear on the lower steeper part of the northern slope of Brigham Hill.

On the northwest side of Brigham Hill, a long slate tongue extends southward along a ravine to the northern foot of the secondary summit where it passes beneath the West Sutton. This slate is mapped as the Parker because at least three exposures of Dunham form anticlinal inliers within the tongue and several exposures of a quartzitic dolomite similar to the Rugg Brook form synclinal inliers in the tongue. East of the tip of the slate tongue the Dunham extends westerly over the slate and is in contact with a quartzitic dolomite similar to the Rugg Brook. There is apparently a minor thrust here.

High on the western slopes of this slate ravine, the trace of the thrust can be followed through the woods to the north. At the north end of the hill the trace of the thrust shifts westward, and the thrust contact is well exposed in the woods part way down the northern slope. Here the competent West Sutton rests upon the incompetent Parker; the Parker is much crumpled, and there is a gouge zone at the contact a few inches thick. The thrust plane is irregular but dips gently west or northwest. From here the thrust roughly follows the contours of the topography, although it cannot be traced precisely until one reaches a point directly north of the eastern edge of Colchester Pond. It can then easily be traced northward; its contact is well exposed in many places. In the vicinity of the former site of School No. 3, the fault plane is very steep. Breccias like those of the Rockledge crop out over a considerable area immediately east of the thrust. The breccia lies on the slate which forms the base of the Parker slate tongue. The pattern of the Brigham Hill thrust in this locality indicates an essentially flat but locally steeply folded and somewhat thin, almost completely eroded westerly portion of the Brigham Hill thrust slice. Less than a mile north of the school site, the fault swings sharply southward to Colchester Pond and then follows the foot of the Oak Hill escarpment to the Winooski Valley; it is bordered on the west by slates of the Rosenberg succession. In the Winooski valley the trace of the thrust passes easterly into a deep re-entrant where the Gilman, Dunham, Parker, and Rugg Brook (?) formations of the Brigham Hill slice are truncated on the southwest by the fault which is bordered in that direction by slates and carbonate rocks of the Rosenberg succession (Cady, 1945, Pl. 10). Near Williston the trace of the Brigham Hill thrust apparently passes beneath the Hinesburg thrust slice (Cady, 1945, Pl. 10, p. 567–568).

Hinesburg thrust.—South of the Winooski River a little northwest of the village of Williston, the trace of the Hinesburg thrust passes

around the west and southwest sides of a hill (elevation 680 feet) formed on the West Sutton and/or Pinnacle formations; the Gilman, Dunham, Parker, and Rugg Brook formations of the adjacent footwall block crop out in parallel bands that strike about N. 10° E. directly into the thrust trace along the southwest side of the hill. Southward these formations strike into the Danby formation of the Rosenberg succession (Cady, 1945, Pl. 10). Therefore, they are bounded in this direction also by a thrust which is here mapped as the Brigham Hill thrust. The Hinesburg thrust follows the Hinesburg escarpment in a swing to the west, turns sharply southward for a few miles, and then angles off to the southeast, where it either dies out or cuts into the as yet undifferentiated rocks of the Green Mountains. A good exposure of the thrust is seen about 1¼ miles north of Hinesburg and a little less than a mile north-northwest of Mechanicsville. Here the Gilman quartzite (the writer's identification) rests upon the Shelburne marble (Cady, 1945, Pl. 10). The thrust plane is almost horizontal.

Possibly the Hinesburg thrust is the southward continuation of the Brigham Hill thrust. In the area north of Williston, the traces of the two thrusts either cross, or one is the continuation of the other—in which case another thrust passes beneath both the Brigham Hill and the Hinesburg thrust slices. Because of lack of evidence proving continuity, they are treated as separate thrusts. The Brigham Hill thrust is considered older than the Hinesburg because its trace apparently passes beneath the latter thrust slice.

Arrowhead thrust.—The Arrowhead thrust is excellently exposed high on the west side of Arrowhead Mountain about 2½ miles north of Milton Village. Its trace may be followed from about a mile north of the mountain to a short distance south of the lower powerhouse on the Lamoille River west of Milton Village. Both to the north and south the trace is lost in slate terranes; a fault in them could be traced only with difficulty. The Gilman quartzite forms the front of the thrust plate from Arrowhead Mountain to the Lamoille. Both north of the mountain and in the river bed the Dunham forms its front. The slates of the Rosenberg succession everywhere form the footwall block along its known trace.

On the north, west, and south sides of Arrowhead Mountain, the contact between the cap of Gilman quartzite and the underlying slates may be observed high up on the slope, where the quartzite forms a nearly vertical cliff. The slate is much contorted, and the Gilman is shattered in places. On the south slope blocks of dolomite occur along and apparently beneath the contact. The cap of Gilman dips eastward which carries the Gilman down beneath the reservoir; it reappears eastward to form the Oak Hill escarpment. Between the reservoir and the escarpment the Gilman crops out in anticlinal inliers in the overlying Dunham. Thus the structure here is a rather broad relatively open syncline with minor wrinkles.

The thrust is excellently exposed at the lower powerhouse in the village of Milton where the Lamoille River has cut a gorge through the thrust plate, here about 100 feet thick. Quartzitic limy slates of the Rosenberg succession that are well stratified and remarkable for their lack of distortion even where blocks of it have been caught up between small thrust slices lie beneath the thrust plane. The Dunham dolomite forms the overthrust block in the gorge and for a short distance south of it; the Gilman outcrops on the north side of the gorge and can be traced almost continuously to Arrowhead Mountain. Along the north side of the gorge the Gilman appears to overlie the Dunham; no contacts are visible. The structure may be explained in part by plunging of the Gilman steeply southward and in part by a possible minor thrust between it and the Dunham. The main thrust dips eastward at 5°–10°.

Arrowhead Mountain has been designated a klippe by Jacobs (1938, p. 23). The structure in the Milton-Arrowhead region is partially obscured by surficial deposits along the base of the Oak Hill escarpment and in the lowland to the west, except in the immediate vicinity of Milton village where outcrops of Dunham are abundant. There is no evidence of a thrust at the base of the escarpment. Thus the only thrust to which a klippe could be genetically related and which has the necessary displacement is the Brigham Hill thrust east of Georgia Mountain, the highest point along the Oak Hill escarpment between the international boundary and the Winooski River. Considering

the behavior of the Brigham Hill thrust on Brigham Hill, the possibility that this thrust block moved over Georgia Mountain cannot be overlooked. However, until the apparently unbroken stratigraphic succession from Arrowhead Mountain to the escarpment and beyond is disproved, it is the writer's opinion that the Milton-Arrowhead rocks do not form a klippe, despite the analogy of the distribution pattern of all other isolated hills in the Champlain Lowland north of the Winooski.

Fairfield Pond thrust.—The fault is named from its exposures along the Oak Hill escarpment 1½ miles west of Fairfield Pond and a little east of Greens Corners. The plane of this thrust truncates a rather extensive though secondary anticline in the younger strata of the homoclinal succession along the Oak Hill escarpment; the thrust slice partially overlies the northern portion of the anticline, and in this area an isolated remnant of the slice may form the Bridgeman Hill klippe at the extreme eastern edge of the Champlain Lowland. Farther south in the region of St. Albans, other isolated remnants of the slice may form the St. Albans Hill and Aldis Hill klippen.

The fault extends southward from the international boundary across Minster Hill and parallels the Oak Hill escarpment to the latitude of Greens Corners where it breaks through to the base of the escarpment, which it follows to the western side of Bellevue Hill southeast of St. Albans. From here its course is uncertain. The thrust zone is clearly visible in the type locality; it can best be seen on the southwest slope of the hill which forms the escarpment north of the road from Greens Corners to Fairfield Pond. Here it is formed on the east limb of the anticline mentioned earlier. The dolomite at the base of the hill is sheared so as to resemble a slate, whereas the dolomite out in the Lowland is unaffected. Farther up the hill begins a series of small slices that may be observed all the way to the top. The Gilman is shattered in places and sheared in the less massive facies. South of Fairfield Pond, the thrust plate is eroded for some distance to the east, forms a prominent re-entrant, and demonstrates the relatively low dip of the fault plane. The Dunham dolomite of the footwall block is well exposed in the re-entrant, and a little of the Parker slate crops out in the extreme eastern part, apparently in the east limb of the anticline. The fault may be traced southward along the base of the escarpment to the latitude of Aldis Hill. The Dunham, part of the Parker, and presumably higher formations are cut out. There are no exposures along the base of the escarpment between Aldis Hill and Bellevue Hill. Northward from the type locality at the Fairfield Pond road the fault is inferred from shearing effects in the Dunham of the footwall block for at least 2 miles; these effects may be related to an abrupt change to a more steeply dipping fault plane north of the re-entrant at the type locality. Farther north toward the Missisquoi River, tracing the fault becomes more difficult. North of the Missisquoi the Dunham appears to be partially if not wholly cut out for a short distance as the fault cuts westerly across it and follows the western edge of the Dunham belt instead of the eastern. In this vicinity the topography is lower east of the fault than to the west, suggesting that the fault plane dips steeply. Along the valley west of Rice Hill, the Dunham outcrops within a few feet of the fine-grained argillaceous phase of the Gilman rather than with one of the two upper phases of it. In two places, the Gilman is brecciated near the contact. Three miles north of the Missisquoi, the trace of the fault appears to cut across the Gilman belt at the axis of the subjacent anticline to the eastern edge of the Champlain Lowland, where, for a short distance, the fault partially, if not wholly, cuts out the Dunham of the west limb of the anticline, for the overthrust Gilman crops out within a few yards of the Parker (?) slate on the west limb of the anticline. From here it follows northeastward to the Minister Hill region near the international boundary, where the Gilman at the axis of the subjacent anticline is broadly exposed. The thrust is clearly indicated a little south of Minster Hill where a thin wedge of Parker slate is in contact with the Gilman on the west and with the Dunham on the east. The fault is probably responsible for the double summit of Minster Hill.

South of Bellevue Hill the course of the fault is uncertain. Southward there is no evidence of a thrust along the base of the escarpment for 12 miles. For the first 3 or 4 miles, exposures of Dunham may be traced from the longitude of Oakland to the Gilman which forms the top

of the escarpment. The trace of the thrust could swing westerly and die out in the slate terrane west of Oakland; if so, the St. Albans klippe could not be genetically related to it. The Fairfield Pond thrust could connect with the Arrowhead thrust in this same slate terrane. The validity of this interpretation rests largely on the identification of a dolomite slightly less than a mile southwest of Oakland. If this dolomite is the Rugg Brook, there is an apparently unbroken stratigraphic succession from the Skeels Corners slate (which outcrops just west of the dolomite) to the Gilman quartzite at the top of the Oak Hill escarpment, and no fault is necessary to explain the structure. However, such a succession does not necessarily indicate the absence of a thrust, which would here parallel the axis of the major syncline of the Champlain Lowland. If the dolomite is the Dunham, a thrust is indicated, and probably the Arrowhead thrust is the southward continuation of the Fairfield Pond thrust. A less plausible possibility is that the Fairfield Pond thrust swings easterly across the southern end of Bellevue Hill, continues southward, and dies out in the Lamoille River region. That would account for the pronounced decline in the elevation of the Oak Hill escarpment between Bellevue Hill and the Lamoille, and could explain possible small displacements north of the Lamoille and east of Cushman Hill.

Klippen west of the Oak Hill escarpment.—From south to north these klippen are Cobble Hill, St. Albans Hill, Aldis Hill, and Bridgeman Hill. They and their related structures form rather prominent hills on the Champlain Lowland a little west of the Oak Hill escarpment. They lie chiefly above the undifferentiated slates of the Rosenberg succession, and a slate interval commonly separates them from the Dunham dolomite and the Gilman quartzite of the homocline or the west limb of the anticlinorium, the westernmost development of which determines the position of the Oak Hill escarpment. At first glance these klippen appear to lie above a single fault plane that emerges at the extreme foot of the escarpment west of the westernmost exposures of Gilman and Dunham. However, such a single fault cannot be demonstrated, and there are indications that the klippen are erosional remnants of the Brigham Hill and Fairfield Pond thrust slices. The western erosional front of the Brigham Hill thrust slice, where it extends westerly across the southern part of the anticlinorium, lies west of the front of the escarpment as developed to the north and is in line with the distributional pattern of the klippen, which suggests by analogy that the Cobble Hill klippe and possibly the Milton-arrowhead rocks as well as the St. Albans Hill and Aldis Hill klippen may be erosional remnants of the Brigham Hill thrust slice.

A little over 2½ miles south of Milton Village and a little over a mile west of the trace of the Brigham Hill thrust, dolomite like the Dunham is exposed high on the eastern side of Cobble Hill and is therefore mapped as such. It dips 40° E. Slate of the Rosenberg succession underlies the dolomite. The bedding in the slate dips about 35° E. on the west side of the hill but about 70° E. beneath the hill. Thus the contact of the slate and the dolomite is apparently a thrust, and the dolomite of the hill forms a klippe, most probably genetically related to the Brigham Hill thrust.

St. Albans Hill may be considered as consisting of the southern main mass of the hill and the much smaller northern knob. The main hill is Gilman quartzite on slate of the Rosenberg succession. At least one small patch of Dunham dolomite, conformably on the Gilman, may be seen in the woods east of the summit. On the northern knob and in the intervening saddle is a maze of outcrops of Gilman and Dunham, and of slates, sandy dolomites, and breccias whose structures are uncertain because of possible mistaken identity in the succession, Oak Hill or Rosenberg, to which they belong. The Gilman and the Dunham compose the knob, and both rest on slate of the Rosenberg succession. The Gilman is in contact with the slate about 40 feet above the base of this knob. Two small bioherms of massive gray limestone in the slate occur at the foot of this slope. The Dunham extends from the north-northwest slope of the knob in a narrowing band around the western slope at the north end of the saddle where it abruptly ends. The dolomite looks like a band wrapped around the western margin of the Gilman. A few rods east of the knob, a small patch of Dunham is apparently conformable upon the

Gilman. The advancing thrust block of Gilman apparently caught up a block of Dunham and pushed it out ahead of the Gilman until the latter rode up over the Dunham. In the saddle exposures of Gilman quartzite alternate with slate. Slate and dolomitic quartzite crop out on the west side of the hill. The relationships suggest repeated slicing on a very small scale and possibly a "smearing out" of portions of the incompetent formations beneath the thrust block.

The Gilman quartzite crops out between the summit of the hill and the Fairfax-St. Albans road. It disappears beneath cover. Only a few tens of feet of the upper Gilman is exposed on St. Albans Hill. That this is uppermost Gilman is attested by the patches of Dunham on the Gilman and by the unusually slaty Gilman on the east slope, a common character of the transitional zone. Perhaps its slaty character is responsible for the lack of an eastward-facing escarpment along the eastern edge of the klippe. The thrust plane dips 5°–10°. On the hill the dip is easterly, but along the eastern edge of the klippe it may be westerly.

The next outcrops to the east are of Skeels Corners slate, here identifiable by *Bovicornelum vermontense* Howell. Between it and the Gilman to the west, the Dunham, Parker, and Rugg Brook are missing; all three occur in proper order east of the Skeels Corners slate. Despite the cover in the valley west of the road, distribution of outcrops to the south suggests that the formations are cut out by a thrust along the eastern edge of the valley west of the road. St. Albans Hill and Aldis Hill are in the latitude where the Brigham Hill thrust apparently dies out to the north and the Fairfield Pond thrust dies out to the south (unless the Arrowhead thrust is a continuation of it). Thus displacement on either fault may not have been great enough in this vicinity to account for the rocks in these klippen, and thus a thrust closer at hand would have to be called on. This argument favors the interpretation that the Arrowhead thrust is the southward continuation of the Fairfield Pond. However, the Brigham Hill thrust appears to die out to the north in an extensive area of West Sutton slate east of Fairfield Pond, in which a thrust might continue undetected several miles north, thus allowing for sufficient displacement to account for the St. Albans Hill klippe as an erosional outlier of the Brigham Hill thrust. It seems difficult to derive the Aldis Hill klippe from the Brigham Hill slice but more so to derive it from the Fairfield Pond slice if the Arrowhead thrust is not a continuation of the Fairfield Pond.

The Aldis Hill klippe is similar to the St. Albans Hill klippe except that in the former only the Gilman is in contact with the slates of the Rosenberg succession along the front of the thrust. Near the base of the southeast slope of the hill, blocks of Dunham appear between the Gilman and the Dunham, possibly dragged beneath the thrust plane. Small patches of Dunham appear above the Gilman near the top of the hill and on the eastern slope as they do on St. Albans Hill. As already suggested, the distributional pattern of the Aldis Hill and St. Albans Hill klippen and of Arrowhead Mountain, as well as the similarity of the rocks above the thrust plane (*i.e.*, a few tens of feet of uppermost Gilman with patches of Dunham lying on it) and the apparent dragging under of blocks of Dunham in all three cases, suggests strongly that the rocks of all three hills are related to the same thrust. This thrust is the Fairfield Pond- Arrowhead thrust if the Milton-Arrowhead rocks do not form a klippe, but may be the Brigham Hill thrust if they do. The Cobble Hill klippe may be related to the same thrust; this thrust may be the Fairfield Pond-Arrowhead thrust (the main trace of which may underlie the Brigham Hill thrust slice east of Cobble Hill), and the thrust between the rocks of the Oak Hill succession and the Danby formation in the Williston region (previously mentioned in the discussion of the Hinesburg thrust) may be its southward continuation.

The Bridgeman Hill klippe lies close to the trace of the Fairfield Pond thrust and thus may be genetically related to it. The Gilman quartzite forms the bulk of the hill, and a continuous band of Dunham dolomite extends along the western edge. The conformable Gilman-Dunham contact is not far down the western slope on the south end of the hill; it descends to near the base of the hill on the northwest slope, where slate outcrops a few feet from the Gilman with a strike that should

carry it beneath the Gilman within a few tens of feet. Thus the Gilman and the Dunham are thrust upon slates presumably of the Rosenberg succession. Cleavage-bedding relations are obscure and so closely parallel that the stratigraphic position of the slate with respect to the Dunham is inconclusive. Parker slate may form the hanging wall; or slate of the Rosenberg succession may form the footwall, in which case the thrust plane is steeply folded. Westward slate outcrops continuously far out into the lowland, except for a very narrow band of dolomitic quartzite and limestone breccia. The position of the thrust trace on the western side of Bridgeman Hill must await the completion of detailed work in this slate belt.

There are few exposures on the north slope of the hill. The Gilman outcrops high on the slope, slate near the base. Gilman rests on slate on the northeast slope. Thus the Gilman seems to be thrust on the slate around the northern end of the hill. The trace of a thrust can be followed almost continuously along the east side of the hill. At the bottom of the slope a belt of Gilman dips steeply eastward and is separated from the Gilman on the main part of the hill by a strip of Dunham. The western edge of this Dunham adjoins footwall slate in some places and hanging-wall Gilman in others. Within the slate belt at the northeast end of the hill are excellent exposures of limestone breccia (Pl. 2, fig. 3) most like those typical of the Rosenberg succession. The eastern belt of Gilman, the strip of Dunham, the slate, and the limestone breccia all lie in the footwall block. The western outcrops of Gilman are continuous to the top of the hill and beyond and belong in the hanging-wall block, and therefore in the klippe. The Gilman dips west at a low angle on the east side of the klippe; the dip steepens near the top of the hill to about 35°–40°. At one place near the north end of the east slope, hanging-wall Gilman forms a small projection to the east where it is in contact with the Dunham. It overlies the Dunham, and the contact is inferred to be a thrust. The dip contact dips about 20° W. Some outcrops of the footwall Dunham show closely spaced essentially horizontal shear cleavage. The thrust plane appears to be gently rolling to almost flat. The trace of the Fairfield Pond thrust at the base of the escarpment east of Bridgeman Hill must continue for several miles into Quebec if the klippe is an erosional remnant of that thrust.

Minor thrusts.—A minor thrust on the low hill northeast of Bridgeman Hill and west of Minster Hill cuts out the strip of Dunham extending northward from the eastern side of Bridgeman Hill, so that the Gilman rests directly upon slate. Within a quarter of a mile the Dunham comes in again with its normal width of outcrop. Half a mile east of East Highgate another minor thrust cuts out the Dunham for at least a few hundred yards, so that the Gilman rests directly upon slate. Possibly, a tear fault exists at the northern end of the Brigham Hill thrust, for the homoclinal belt of White Brook, which extends southward through Fairfield, suddenly shifts westward about a mile but cannot be traced across this area despite abundant outcrops. Moreover, in this area the coarse-grained Pinnacle abruptly changes northward to fine-grained slates of the West Sutton formation. There is possibly a fault on the southern and western slopes of Cushman Hill (southeast of St. Albans) where the Pinnacle formation rather than the West Sutton may be in contact with the Gilman quartzite. Uncertainty arises from the fact that the West Sutton as well as the Pinnacle contains coarse-grained graywacke facies in western areas, and the White Brook dolomite which commonly separates the two is absent from almost all the Cushman Hill area. The inferred position of the White Brook horizon in this area is mapped (Pl. 1) to make clearer the structural pattern and to avoid leaving isolated occurrences of the White Brook north and south of Cushman Hill. In the valley west of Rice Hill a mile southwest of Franklin Village, a part of the belt of Dunham dolomite is sheared as at the type locality of the Fairfield Pond thrust. This may be due to either a separate minor fault or a branch of the Fairfield Pond thrust. Other minor faults may occur between the Gilman and the Dunham at the lower powerhouse in Milton Village and west of the summit of Brigham Hill.

Mechanics of Deformation

Compressive stresses folded the Oak Hill succession. The necessary elongation parallel

to the fold axes was accommodated in part by the production of a salient whose axis passes through the St. Albans region, and in part by the repeated plunging of the folds, producing depressions and culminations (Billings, 1942). Folding was more profound on the south limb of the salient; possibly greater compressive stresses were conditioned by proximity to the Adirondack massif. Folding progressed until fractures developed and thrusting replaced folding. The initial fractures were, presumably, break thrusts in the more competent strata at or near the anticlinal crests of the folds. With partial release of the confined stresses consequent on formation of the initial fractures, the general zone of fracture deepened and the thrusts extended downward easterly as well as up to the west across the folded structures forming subsequent shear thrusts. All but a few of the sites of initial "breaks" have been eroded from the fronts of the hanging-wall blocks which now cover the footwall sites. The formations chiefly involved are those above the West Sutton slate—the Gilman, Dunham, Parker, Rugg Brook, Skeels Corners, and whatever higher formations were present, and which range in competency from high to low in nearly respective order. Thus the initial fractures probably formed chiefly in the Gilman and the Dunham. Viewed at right angles to the strike of the fold axes, the thrust surfaces thus formed arched in the vicinity of the highest culminations and bowed downward in the vicinity of the lowest depressions.

In the Arrowhead Mountain-Milton Village region, at the lower powerhouse and immediately north of the mountain, the thrust sliced through the Dunham in a structural low (depression). North of the village toward Arrowhead Mountain, both the thrust plane and the Dunham rise, the Dunham more steeply than the thrust; both approach a structural high (culmination) in this direction. Thus the thrust sliced through the Dunham into the upper part of the Gilman. North of the mountain both the thrust plane and the Gilman pass downward, the Gilman more steeply than the thrust plane, so that the thrust plane slices through the Dunham and presumably into younger formations, possibly continuing northward and northeastward in these higher formations. The Cobble Hill, St. Albans Hill, and Aldis Hill klippen probably owe their origin to similar behaviors of thrust planes to structural highs, although the arching of the thrust planes cannot be demonstrated as in the Milton-Arrowhead Mountain region.

Bridgeman Hill has a much greater thickness of Gilman and Dunham exposed in it than do the other klippen. This hill is at the latitude of the homoclinal area of the Oak Hill succession, in that part of the slice characterized by many relatively small doubly plunging folds arranged en echelon but with no major folds. With such a pattern it would be impossible for the upwarps and downwarps of the thrust plane to conform everywhere to the upwarps (culminations) and downwarps (depressions) of the folds. A downwarp in the thrust plane may have sliced through a culmination in the folds.

The events connected with deformation of the rocks of the Oak Hill succession subsequent to initial folding are conceived to be as follows: Relief of the compressive stresses through formation of the St. Albans salient and the lesser culminations and depressions was not great enough to stop further deformation. Consequently, thrusts formed; the hanging-wall blocks moved westerly onto rocks that now underlie the Champlain Lowland. The distance moved was only a few miles and in some cases may have been less than a mile. When thrusting along these faults stopped, further application of the compressive stresses folded the thrust planes and the rocks above and below them on axes roughly coincident in strike and plunge with those of the folds formed previous to thrusting. A notable illustration of this is the southerly plunge of the axis of anticlinal warp of the Brigham Hill thrust plane coincident with the plunge of the axis of the anticlinorium on the south limb of the St. Albans salient.

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Short Notes

TRENDS IN GEOMORPHIC RESEARCH

By Charles W. Carlston

In April 1950 the writer spoke before the geology club of a midwestern university on trends of research in geomorphology. To obtain quantitative data on types of research done in the general field of geomorphology during the past 20 years, titles and abstracts of papers given at The Geological Society of America meetings were analyzed. Proceedings and Programs of the Society provide a convenient source for these data. Virtually all geomorphologists try to report on their work at meetings of the Society, and their papers provide a timely summary of completed research since papers given at meetings of the Society usually are concerned with research completed during the preceding year. The full publication may not appear for several years after completion of research. Results of the analysis of geomorphic papers presented at meetings of the Society from 1931 to 1940 and 1945 to 1949 are shown on Figure 1. The meetings are those of the parent Society, Cordilleran Section, Rocky Mountain Section, and Section E, AAAS. It is thought that the results will be of interest to geologists, particularly those in the field of geomorphology.

The papers were classified into four general categories:

(1) Glacial Geology: Studies of glacial processes, deposits, erosional forms, including research on Pleistocene stratigraphy, cryopedology, but not including studies of marine terraces and loess deposits beyond the limit of Pleistocene glaciation.

(2) Basic Studies: Papers on research outside the field of glacial geology whose primary object is the investigation of some geomorphic process, agency, type of land form. An example: *Frequency-distribution analysis of slopes*, by A. N. Strahler, Geol. Soc. Am., Bull., vol. 60, p. 1975 (1949).

(3) Eastern U.S.: Papers on the results of geomorphic studies of regional problems in the eastern United States and eastern Canada, outside the field of glacial geology. An example: *Blue Ridge—a fault scarp*, by W. A. White, Geol. Soc. Am., Bull., vol. 57, p. 1242 (1946).

(4) Western U.S.: Papers on the results of geomorphic studies of regional problems in the western United States and western Canada, outside the field of glacial geology. An example: *Dune forms in western Nebraska*, by H. T. U. Smith, Geol. Soc. Am., Bull., vol. 60, p. 1920 (1949).

A fifth category, geomorphic studies outside of North America, was included in the computations but is not shown on the graph because of the very few papers in this group.

The graph shows a decrease in total number of papers in the whole field of geomorphology in the three years preceding the war. Productivity in the field has gone upward in post-war years, but not to the level of the early 1930's. Research in glacial geology increased in the late 1930's in terms of percentage of papers, and that increase has continued in post-war years to more than half the total number of papers in 1945, 1946, and 1949. The trend in volume of research in glacial geology lends considerable support to those glacial geologists who believe that their field should no longer be considered a section of geomorphology, but should be recognized as a separate branch of geology.

The number and percentage of geomorphic papers outside the field of glacial geology has declined in post-war years. There is some suggestion that in post-war years geomorphologists are turning from the east to the western states for research outside of glacial geology, but this trend is of too short a duration to indicate whether it is temporary or significant. Post-war research on basic studies has not declined significantly in comparison with other types of

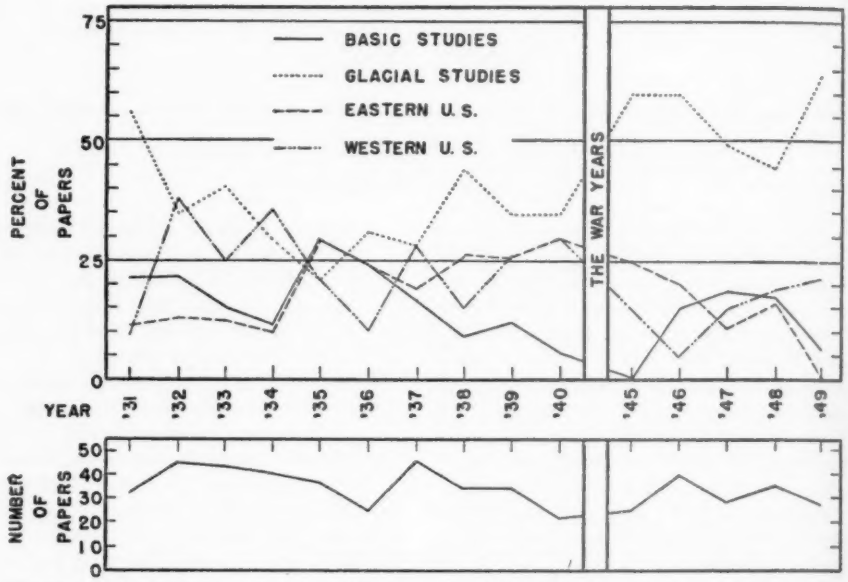

FIGURE 1.—ANALYSIS OF PAPERS ON GEOMORPHOLOGY GIVEN AT MEETINGS OF THE GEOLOGICAL SOCIETY OF AMERICA, 1931 TO 1940, 1945 TO 1949

geomorphic research. A recent development of quantitative research in geomorphology, based on Horton's (1945) quantitative approach to geomorphic study, promises, however, to result in an increase in number of basic studies and in a general increase in geomorphic research.

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POLLEN ANALYSES IN OLD AND NEW MEXICO

PAUL B. SEARS

During the spring of 1948, through the encouragement of Mexican colleagues, the writer began a study of fossil pollen in lake sediments within the Basin of Mexico. Attention was first concentrated on the upper, archeological layers in a search for possible relations between human activity and climatic history. Presently, however, series of samples extending to great depths (60 meters or more) became available. These have since been supplemented by series of comparable length from northern New Mexico. Since sediments from both regions contain an abundance of pollen except at levels where vulcanism or exposure has interfered, they afford a remarkably long and continuous record of vegetation and climate.

Both regions are alike in showing a definite alternation of humid and dry conditions. In number, position, and intensity these alternations strongly suggest a connection with the known episodes of glacial retreat and readvance, which the writer (Ecol., vol. 29, p. 326–333, 1948) believes to be recorded in pollen profiles from northeastern North America.

Of archeological interest are the following points:

(1) The Archaic and Toltec-Aztec cultures were beneficiaries of well-marked moist intervals, favorable to the formation of rich organic soil.

(2) These two great waves of culture were separated by a long dry interval.

(3) Maize pollen in abundance occurs to a depth of 6 meters under the present city of Mexico.

Numerous obligations will be acknowledged when detailed results are published. Thanks should be expressed at this time, however, to my associate, Mrs. Katherine Clisby, through whose skill and persistence numerous technical difficulties were overcome.

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